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# A Numerical Simulation Analysis of Regional Groundwater Flow for Basin Management : Plains Regions, Alberta

by

© Duke Urhobo Ophori

A THESIS
SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH
IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE
OF Doctor of Philosophy

Department of Geology

EDMONTON, ALBERTA
Spring, 1986

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R A Freeze

External Examiner

Date March 27, 1986

# Dedication

Dedicated to my brother Smart

Where are you ?

# ABSTRACT

It is possible to describe the hydrogeologic setting of the Plains Regions of Alberta despite its complexity. This objective was achieved by categorizing the component drainage basins through the concept of the hydrogeologic environment. A baseflow analysis of 32 selected basins served the purpose of a hydrologic characterization for developing a basic understanding of the regional groundwater distribution and movement.

Plots of specific baseflow rate versus basin area [q(A)], display a regularity. This regularity suggests an initial high rate of increase of baseflow rate (q) with basin area (A), followed by a lower rate of increase at larger basin areas.

Numerical-model simulations of two-dimensional vertical sections were run in an attempt to explain the observed regularity by the theory of regional gravity flow of groundwater.

This thesis demonstrates for the first time, the regular distribution of steady-state regional groundwater discharge and its relation to basin area in the Plains Regions of Alberta. Regional groundwater discharge is not haphazard or random but proceeds in a systematic way. Quantitatively, the regularity types indicate greater discharge in regional discharge areas (areas of low topography and including local recharge regions) than in regional recharge areas (areas of high topography and including local discharge regions).

Consequently, unit basin models of the Plains Regions were developed systematically by numerical simulation to establish the principles governing the relations between well yield and well location, and between well location and basin stability during development. These relations are expressed in terms of transitional-basin-yield (TBY), and sustainable-basin-yield (SBY). TBY is defined as the cumulative inflow into the system during and due to development at a particular site from the initial to the final steady-state conditions. The SBY, on the other hand, is the amount of water captured from precipitation due to development at a particular site at the newly established steady-state conditions.

In general, TBY is highest if wells are located in the discharge area, and it decreases gradually with the wells in the recharge area. On the other hand, SBY is greater with wells in the discharge area than with the wells in the recharge area. From a basin hydrological viewpoint, these are key relations to planning the optimal location of development sites in a basin. In the location of a site of development in a regional setting, a consideration of capture from atmospheric moisture may be as important as considering the capture of natural discharge.

Through these models it has been possible to analyze the stability of a basin under development quantitatively. It is shown that under conditions of restricted rainfall, a basin can be developed on a sustained basis by locating a well in the discharge area, whereas groundwater mining results in the same basin under the same conditions, but with the well in the recharge area. On this basis, regional groundwater exploitation should be initiated in discharge areas and moved only gradually away from it for compelling reasons.

Several factors influence TBY, SBY, and basin stability to various degrees. These factors include well location, well depth, permissible proportion of available drawdown, precipitation rates, regional slopes, basin depth, depth to the water table, hydraulic conductivities, ratios of hydraulic conductivity to precipitation rates, and the positions and sizes of aquifers. Since the effects of these factors can be pre-determined before development takes place, their analysis may guide in planning the optimal exploitation of regional groundwater resources.

An application of the models to groundwater management in Ross Creek Basin shows that groundwater pumpage is insignificant relative to demands. Some areas in the basin should be aided artificially to enhance natural groundwater discharge from springs. Licenses may be issued for the development of springs for local use. Predictions of groundwater yield based exclusively on aquifer hydraulics may be different when basin hydrology is considered. Aquifer hydraulics does not provide the response of the entire groundwater regime to development.

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# 1. INTRODUCTION

## 1.1 Background of the study

Water is one of the most abundant natural resources on the earth. This fact often leads to a common misconception that water is an inexhaustible resource. However, this misconception is being dispelled rapidly by the ever-increasing awareness that water is only a renewable substance which is exhaustible.

In addition to the limited occurrence of water, increases in population, and in agricultural and industrial activities result in greater demands. The ultimate solution to expanding needs is the expansion and improved management of existing facilities, and/or the development of additional supply systems. In either case, the goal is to obtain an adequate quantity of water that also meets predetermined quality requirements at least cost.

The Plains Regions of Alberta occupy areas of geographic imbalance between greatest water supply and greatest water demand. The southern Plains have about 15% of the total water supply and 80% of the total water demand (Alberta Environment, 1983). This imbalance, together with rapid economic development, intensive urbanization, and increased standards of living, have resulted in high demands for water in this area. Consequently, a water management philosophy is in force, part of which is based on "better use of available water resources", the basic premise on which this thesis is developed.

In the past, large amounts of money and research efforts have been expended in the planning and management of surface water supply systems. Despite the enormous volumes of flow associated with surface water reservoirs, some of these early research ventures pointed toward the need for supplementary water sources.

Most groundwater management studies have been prompted by reasons of supplementing surface water supplies for quantity and quality control, regulating seasonal fluctuation of surface water supplies, or storing flood water through injection followed by release and use in the time of need. In few cases, groundwater is the primary source of water supply, and aquifer management becomes the tool for optimal allocation and preservation of groundwater, whereas, available surface water is a source of supplementary supply. Often, aquifer management focusses on only the high permeability units of the groundwater basin, disregarding that the basin is a large natural underground reservoir, and that utilization of groundwater by one user affects the water supply of nearby users. An optimal development of groundwater resources must be planned in terms of the entire groundwater basin. An understanding of the inter-connection between groundwater flow paths in the entire basin is necessary in planning an optimal operational strategy for its exploitation.

Several studies have been conducted in this regard. Such studies, concerned with individual or general principles and concepts of regional groundwater flow have resulted in a great deal of information on groundwater movement, flow pattern types and distribution, groundwater chemistry and distribution, and their controlling parameters within the basin. However, the transfer value of the information with respect to groundwater quantity management within the basin is limited, mostly because the studies have not been aimed particularly at this objective.

A high percentage of research efforts to quantify regional groundwater flow have assumed simplified models according to which drawdown due to development is distributed evenly across the basin. This assumption suggests that the same degree of development in any part of the basin will result in the same amount of groundwater yield across the basin. The degree of development is measured by a given permissible drawdown, whereas the amount of groundwater yield may mean either of two things: 1. the quantity of water that is induced to flow through the basin by development at a particular site; 2. the rate at which water can be captured into the groundwater system from atmospheric moisture (eg. precipitation) by development at a site. The practice of simplified models can lead to a serious misunderstanding about regional groundwater yield, especially in heterogeneous basins. Furthermore, even in homogeneous basins, our present knowledge about flow pattern distribution indicates that groundwater yields from the same level of development may vary spatially within the basin. Few studies have reported this phenomenon and have often expressed their surprise at its occurrence. Recent trends in research show that increasing emphases are being laid on groundwater development on a regional scale relative to the local aquifer scale in the Plains Regions of Alberta, and in other parts of the world. Only by thorough understanding of the interrelationships of groundwater development, spatial location of development, and groundwater yield can progress be made in alleviating some problems associated with basin planning and management.

Although most regional groundwater studies have involved a consideration of all the components of the hydrologic cycle, atmospheric water (precipitation and evapotranspiration) has been quantified in terms of recharge into the system. Recharge is either determined from observation well hydrographs, estimated from streamflow records, or calculated from mathematical models of groundwater systems. In each case, due to the lack of adequate data, the estimated recharge values may be inaccurate, but they are assumed to be known a priori and treated as such in further analysis of the groundwater system. Yet, in some other studies, recharge is not considered during the evaluation of safe production rates. Apparently, some error is associated with yields calculated without the consideration of recharge, or with yields estimated with the assumption of constant recharge. Variation in recharge that may occur in basins under development is a topic of current research.

It is believed in this study that the science of surface and groundwater management has reached an advanced stage, and that much information exists in the field of regional groundwater flow. However, this information is either incomplete or has not been arranged in an acceptable order for sound regional groundwater management. It would be desirable if one could evaluate and arrange existing and new information on regional groundwater flow in a logical manner for the purpose of basin planning and management. Such an attempt will be useful not only because it may lead to new hypotheses and stimulate further investigations into the subject but may also result in new methodologies for managing regional groundwater.

## 1.2 Literature review

### 1.2.1 Regional groundwater flow

Basic concepts such as Darcy's law, hydraulic potential, hydraulic head, equation of continuity, general equation of groundwater flow, and boundary conditions must be understood in order to evaluate the movement of water through porous media. These concepts are reviewed herein, together with the relevant literature in a form that will highlight the objectives of this study.

One of the earliest fundamental laws of groundwater hydraulics is Darcy's law. It is an empirical force-flux law which states that the flow of water through a saturated column of soil is proportional to the head difference across the column and inversely proportional to the length of the column. For flow in the x-direction of an x, y, z coordinate system in a homogeneous isotopic medium, Darcy's law may be written as:

$$q_x = -K\frac{\partial h}{\partial x} \qquad (1)$$

where: $q_x$ = specific discharge in x direction (m/s);
$x$ = distance in x direction (m);
$h$ = hydraulic head (m);
$K$ = hydraulic conductivity (m/s);
$\frac{\partial h}{\partial x}$ = hydraulic head gradient (unitless).

For a two-dimensional vertical section, a similar expression can be written for the z-direction, and the hydraulic conductivity written $K(x,z)$ if the medium is anisotropic.

The physical meaning of hydraulic head in (1) is discussed by Hubbert (1940), who described the groundwater system as a region at every point of which there corresponds a value of a physical quantity. According to Hubbert (1940), hydraulic potential is the physical quantity in groundwater flow fields, and it is capable of measurement at every point in the field. It has properties such that flow is always from regions of high to regions of low values. Employing the generalized form of the Bernoulli's Theorem, Hubbert related hydraulic potential to the elevation, pressure, and velocity along a given flow line of a fluid. Recognizing that velocities of groundwater flow are negligible, the hydraulic potential in terms of gravitational and pressure potentials is (Hubbert, 1940):

$$\Phi = gz + \frac{p - p_o}{\rho} \qquad (2)$$

where: $\Phi$ = hydraulic potential at any point in the field ($m^2/s^2$);
$g$ = acceleration due to gravity ($m/s^2$);
$z$ = elevation of the point above a standard horizontal datum (m);
$p$ = pressure at the point ($N/m^2$);
$p_o$ = atmospheric pressure ($N/m^2$);
$\rho$ = density of water ($kg/m^3$).

Further simplification reduces (2) to (Hubbert, 1940):

$$\phi = gh \qquad (3)$$

where: h = height of a liquid column above a standard horizontal datum (m).

Hydraulic head is defined as:

$$h = \frac{\Phi}{g} \qquad (4)$$

Equation (4) shows that hydraulic head is also a potential quantity and should obey the laws of potential theory, since it is the hydraulic potential divided by a constant (acceleration due to gravity). Hydraulic head is obtained easily from measuring water levels in piezometers or wells, and it is the quantity that is employed most often in the evaluation of groundwater flow fields.

The equation of continuity is the mathematical expression of the law of conservation of mass which states that the net rate of fluid mass flow into any small elemental volume of matter is equal to the time rate of change of fluid mass storage in the element. The equation of continuity takes the form (Freeze and Cherry, 1979):

$$\frac{\partial(\rho q_x)}{\partial x} + \frac{\partial(\rho q_y)}{\partial y} + \frac{\partial(\rho q_z)}{\partial z} = - \frac{\partial(\rho n)}{\partial t} \qquad (5)$$

where: $q_x$ , $q_y$ , $q_z$ = specific discharge in the three coordinate directions (m/s);

n = porosity of porous medium (unitless);

t = time (s).

A combination of (1) and (5) followed by some mathematical manipulations for a groundwater system in three-dimensions yields

$$\frac{\partial}{\partial x}\left(K_x \frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_y \frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_z \frac{\partial h}{\partial z}\right) = S_s \frac{\partial h}{\partial t} \qquad (6)$$

where: $K_x$, $K_y$, $K_z$ = hydraulic conductivity in the three coordinate directions (m/s);

Ss = specific storage of porous medium ($m^{-1}$); and other parameters are as defined previously.

Equation (6) is the general equation of transient groundwater flow through a saturated anisotropic porous medium. For flow of an incompressible fluid through a rigid non-homogeneous porous medium, (6) becomes the Richard's equation:

$$\frac{\partial}{\partial x}\left(K_x\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_y\frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_z\frac{\partial h}{\partial z}\right) = 0 \qquad (7)$$

or, in two dimensions:

$$\frac{\partial}{\partial x}\left(K_x\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial z}\left(K_z\frac{\partial h}{\partial z}\right) = 0 \qquad (8)$$

If the medium is homogeneous and isotropic, equation (7) reduces to Laplace's equation:

$$\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} + \frac{\partial^2 h}{\partial z^2} = 0 \qquad (9)$$

and, in two dimensions:

$$\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial z^2} = 0 \qquad (10)$$

Equations (7) through (10) are referred to as the equations of flow for steady-state flow through porous media, and have been reviewed in this section because of their use in later sections of this study.

The solutions of (6) through (10) require that boundary conditions for the region of interest be defined. Boundary conditions are mathematical representations of the boundaries of the region of interest, which simulate the effects of materials adjacent to, but outside the region, in the solution of the problem. Boundary conditions constrain the problem, and make solution to the problem unique.

Using various boundary conditions, several forms of (6) through (10) have been solved by different scientists in two-dimensional vertical sections that are relevant to this study.

Hubbert (1940) was the first to concern himself with the large-scale regional effects of his theory of groundwater motion. Qualitatively, he determined the theoretical distribution of flow and equipotential lines in a groundwater system. Figure 1 shows Hubbert's model of flow distribution between two valleys. Figure 1 portrays a large area in which groundwater movement is downward into the system from the water table, and little areas around the creeks where groundwater moves out of the system into the creeks naturally.

Tóth (1962, 1963) introduced the method of mathematical modeling to determine groundwater flow patterns from mathematical solutions to boundary value problems. Figure 2a is an illustration of the flow pattern of Tóth (1962) for a homogeneous medium whose water table elevation increases linearly from the valley bottom to the water divide. In Figure 2a, the midline separates an area of downward movement of groundwater (recharge area) from an equal area of upward movement (discharge area). For a more complicated water table configuration, Tóth (1963) obtained the flow pattern displayed in Figure 2b. In addition to the recognition of recharge and discharge areas, Tóth (1963) defined three types of groundwater flow systems: 1. a local system with a recharge area at a topographic high and a discharge area at an adjacent topographic low; 2. an intermediate system with one or more topographic highs between the recharge and discharge area; and 3. a regional system where the recharge area occupies the basin water divide and the discharge area lies at the bottom of the basin.

It is observed that the greatest density of flow lines lies at shallow depths of local systems (Figure 2b), whereas the density decreases with depth and with transition from local through intermediate to regional flow systems. The intensity of flow through the groundwater basin, therefore, decreases from shallow to deeper zones, and from local to regional systems. The higher the relief, the stronger and deeper are the local flow systems. Increase in slope of the valley flank, increases lateral flow towards the bottom of the valley, degenerates local flow systems, and enhances intermediate and regional flow systems. The existence of high permeability layers may strengthen intermediate or regional systems relative to local systems, depending on their sizes and areal position within the basin. All these factors contribute to a complicated distribution of steady-state regional groundwater flow through a basin.

Freeze (1966), and Freeze and Witherspoon (1966, 1967, 1968) extended the work of Tóth to include generalized solutions covering a wider spectrum of existing field situations. Through the development of numerical mathematical models, Freeze (1966), and Freeze

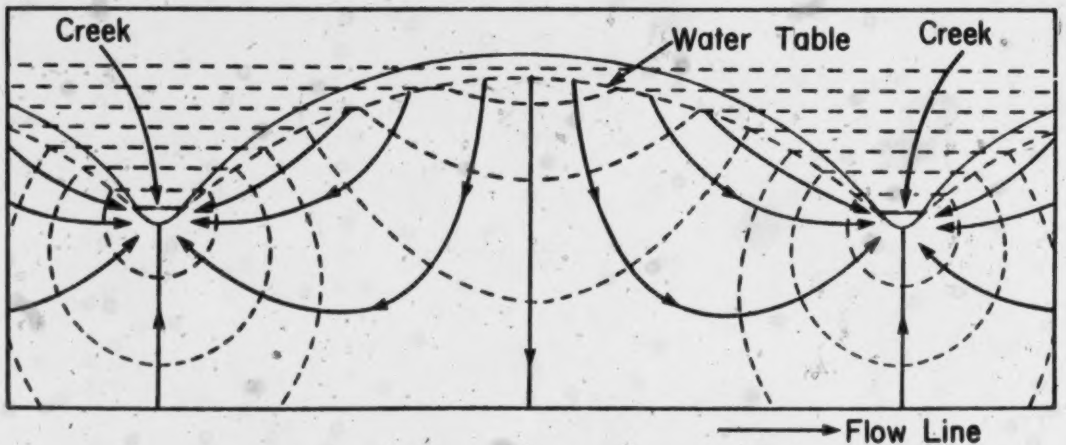

Creek
Water Table
Creek
Flow Line

Figure 1. Schematic flow pattern in uniformly permeable materials between two effluent streams (after Hubbert, 1940).

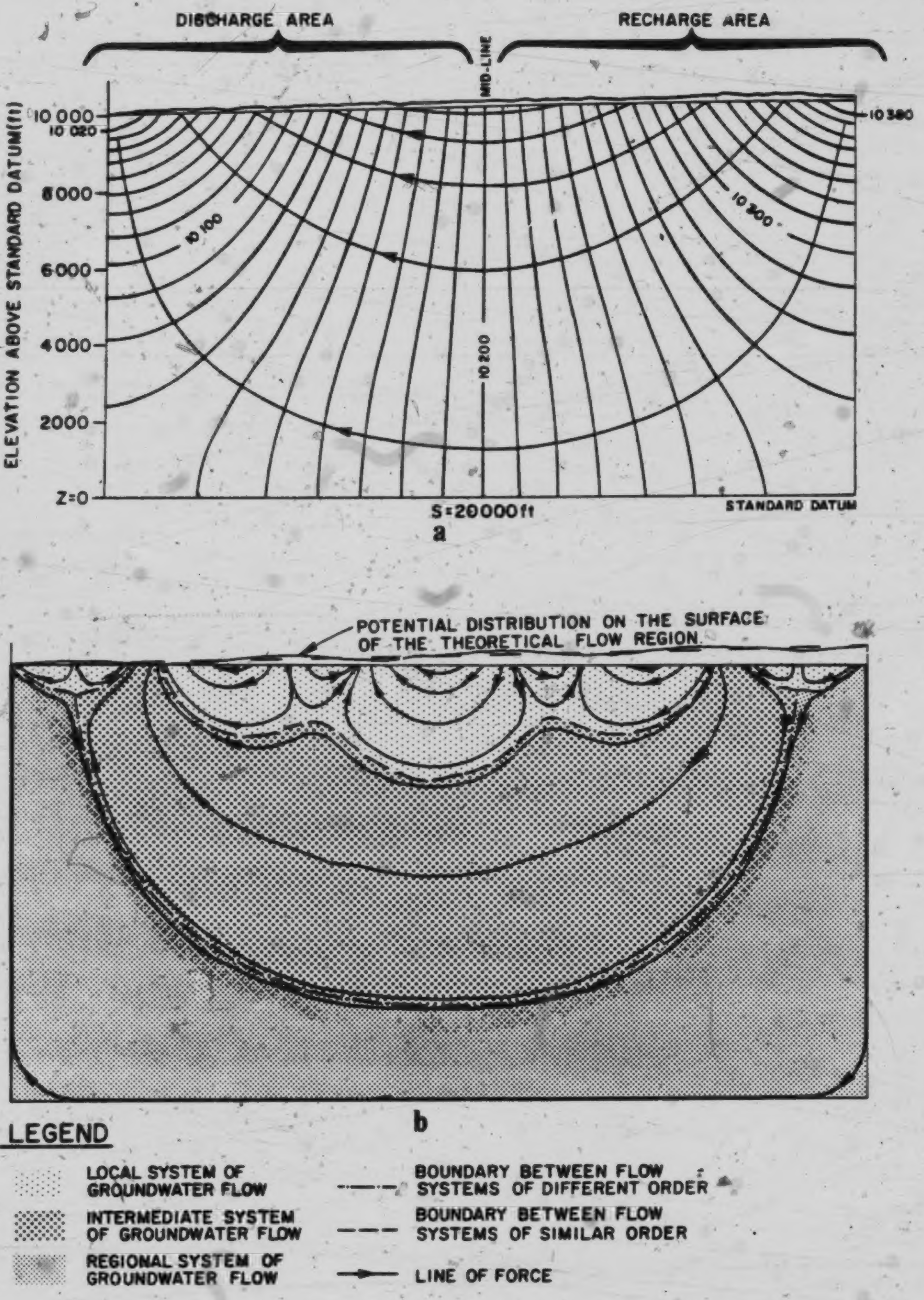

DISCHARGE AREA
RECHARGE AREA
MID-LINE
ELEVATION ABOVE STANDARD DATUM(ft)
10 000
10 020
8000
6000
4000
2000
Z=0
10 100
10 200
10 300
10 380
S=20000ft
STANDARD DATUM
a
POTENTIAL DISTRIBUTION ON THE SURFACE OF THE THEORETICAL FLOW REGION
b
LEGEND
LOCAL SYSTEM OF GROUNDWATER FLOW
INTERMEDIATE SYSTEM OF GROUNDWATER FLOW
REGIONAL SYSTEM OF GROUNDWATER FLOW
BOUNDARY BETWEEN FLOW SYSTEMS OF DIFFERENT ORDER
BOUNDARY BETWEEN FLOW SYSTEMS OF SIMILAR ORDER
LINE OF FORCE

Figure 2. Flow patterns for simple and complicated water table configurations (after Toth, 1962, 1963).

and Witherspoon (1966, 1967, 1968) determined theoretical flow patterns for two- and three-dimensional, non-homogeneous, anisotropic groundwater basins with any water table configuration. Their numerical solutions compared favourably with the analytical solutions of Tóth (1962, 1963), and in most cases, their conclusions have been similar.

Several other studies conducted under the same framework as those reported above have often produced similar and related conclusions regarding the distribution of natural groundwater flow and flow patterns in small and large basins. These studies include those of Fayers and Sheldon (1962), Freeze (1968, 1969a), Parsons (1970), Haefeli (1970, 1972), to mention a few.

### 1.2.2 Basin yield

As opposed to the local aquifer scale of hydrological studies which became popular as early as the mid- and late-1930's, a larger scale approach to groundwater hydrological studies was triggered by Hubbert (1940). In the early 1960's mathematical modeling techniques (Tóth, 1962), and field techniques (Meyboom, 1963) were now being employed in the analysis of regional groundwater flow. The primary objective of these studies was to understand the movement of groundwater, to determine basin yield, and to plan the best use for groundwater.

Freeze and Cherry (1979) defined basin yield as the maximum rate of withdrawal that can be sustained by the complete hydrologic system in a groundwater basin without causing unacceptable declines in hydraulic head in the system or causing unacceptable changes to any other component of the hydrologic cycle in the basin.

Freeze (1966, 1968, 1969b) referred to the quantity of flow through an undeveloped basin under natural conditions as the "natural basin yield". Tóth (1968) believes that Freeze's natural basin yield is the lower limit of the "basin safe yield" since withdrawal in any part of the basin will cause greater recharge into the system than would enter under natural conditions. Similar concepts are also known to exist in the Russian and Hungarian literature, and have been termed "dynamic reserves of groundwater" and "artesian water circulation" respectively. The existence of a limit to the utilization of regional groundwater resources has been noted also by Kovács (1981) who used a characteristic curve of groundwater balance to determine the level of water table below which groundwater resources become unstable.

Tóth (1968, 1982) used a field technique, employing observation well data to estimate the basin yield of a groundwater system in the Plains Regions of Alberta. The estimated basin yield is an average over the entire region, hence no recharge-discharge correlations were made.

Freeze (1971) classified simulation models of basin wide water resource into two groups. The first group (Tyson and Weber, 1964; Bittinger et al., 1967; Pinder and Bredehoeft, 1968), considered transient flow in a two-dimensional horizontal aquifer, but no recharge-discharge properties of the basin were considered. The second group (Fayers and Sheldon, 1962; Tóth, 1963; Freeze and Witherspoon, 1966), considered steady-state flow in two-dimensional, vertical cross sections. Whereas this group gives a good understanding of the recharge-discharge properties of the basin, it does not consider the effects of development. Consequently, Freeze (1971) developed a versatile three-dimensional model to investigate both the effects of development, and the changes in rates of groundwater recharge and discharge induced by the development of a basin. Although Freeze (1971) related the maximum stable basin yield to the recharge-discharge regimes and water level drawdown in the system successfully, no attempt was made to evaluate possible differences that might occur if the point of development were varied spatially throughout the basin.

Several other studies have also considered the "capture" of recharge and/or discharge, or loss of storage in the groundwater system due to development. Theis (1940), Brown (1963), Bredehoeft and Young (1970), Lohman (1972), and Bredehoeft et al. (1982), in their individual studies arrived at positive conclusions toward the adjustment in recharge and discharge rates as a result of pumping. In these studies, spatial variation of the point of development was either not considered, or the variation was too specific to reflect the detailed trend across the basin.

A large number of studies associated with basin yield were prompted by dissatisfaction in the use of the basin-yield-related concept of safe yield. Such studies include those of Kazmann (1956, 1972), Todd (1959), Thomas (1951), Buras (1966), Burt (1967), Bear and Levin (1967), Domenico et al. (1968), and Domenico (1972). Few of these studies suggest that the more rigorous concept of optimal yield should be substituted for safe yield.

The preceding review of literature on basin yield raises several points. First, the concept of basin yield appears to be newer and less well understood than similar concepts, such as well yield and aquifer yield, in hydrological studies. Secondly, there is confusion associated with the meaning of basin yield, especially when the definition of basin yield according to Freeze and Cherry (1979) is adopted. The third point and the one most relevant to the present study, is that the relationship between development and changes in the natural and induced recharge and discharge is not understood thoroughly. This situation is worse if the location of development is varied across the basin. This is one of the areas in which this thesis is different from other studies.

### 1.2.3 Groundwater management models

The evaluation of the responses of groundwater systems has resulted in the development of various groundwater management models. Those models either consider the groundwater system alone, or groundwater in conjunction with surface water. A comprehensive list of available models is given by Bachmat et al. (1980). For the purpose of this thesis, groundwater management models may be separated into two categories: simulation models, and simulation-optimization models. Simulation models produce the most desirable management solutions among a variety of alternatives. On the other hand, simulation-optimization models provide solutions that are best or optimal, considering the physical behaviour of the system, together with stated management objectives and constraints.

Using analog and digital computer methods, various forms of equations (6) through (10) in combination with initial conditions, boundary conditions, and aquifer parameters have been solved by different scientists, to provide the basis for groundwater management. Chun et al. (1964), and Tyson and Weber (1964) used the simulation approach in the evaluation of alternative plans for the conjuctive operation of a basin. Trescott et al. (1976) applied the same approach in the determination of well locations that would result in least drawdown in a river-connected aquifer. Based on simulation models, Pinder and Bredehoeft (1968), Pinder (1970), Rushton and Tomlinson (1971), and Prickett and Lonnquist (1971) studied the responses of different groundwater systems to imposed stresses.

Research trends point toward a proliferation of simulation-optimization models in the past decade. An excellent classification of this group of models is presented in Gorelick (1983). Most of these models are deterministic, hence they employ mathematical programming in finding optimal solutions to management problems. Maddock (1972), Aguado (1974), and Aguado and Remson (1974) discovered the methodological framework for optimal hydraulic groundwater management. Since then, researchers such as Schwarz (1976), Alley et al. (1976), Willis and Newman (1976), Haimes (1977), Heidari (1982), Gorelick (1982), and Gorelick and Remson (1982) have analyzed several groundwater systems for optimal water supply and waste management by simulation and optimization. On the regional scale, simulation-optimization models have also been employed in determining optimal operational strategies for conjunctive water systems (Bredehoeft and Young, 1970, 1983; Young and Bredehoeft, 1972; Downing et al., 1974; Chaudry et al., 1974; Morel-Seytoux, 1975; Ehrlich, 1984).

A detailed evaluation of the studies listed in this section reveals that: 1. most of the studies are technique-oriented, hence the methods of solution have been more emphasized than the practical results; 2. but for few cases where the results are applicable generally, others are site-specific; 3. in groundwater management, the simulation-optimization technique is becoming more popular than the simulation technique; 4. none of the studies discussed the consequences of natural regional groundwater flow in a systematic manner, a point of great concern in the present study.

Another difference between the two groups of management models is that simulation models are executed repeatedly under different management scenarios. To this end, it becomes expensive and impracticable to test the numerous possibilities for best results. However, the repeated execution leads to new hypotheses which will otherwise be uncovered. On the other hand, simulation-optimization models provide immediate optimal answers that might not be adequate for further consideration. Simulation models are used in this thesis for this reason.

### 1.3 Problem setting

The works of Tóth (1962, 1963), buttressed by Freeze (1966), and Freeze and Witherspoon (1966, 1967, 1968) gave rise to the relatively new branch of hydrological studies termed "regional groundwater flow".

In this study, the most important concepts in regional groundwater flow that will keep recurring are those of the existence of a recharge area, and a discharge area separated by a midline (Figure 3). Figure 3 shows that under natural steady-state conditions, water moves through the system in the following manner: into the system in the recharge area (point A), away from the recharge area inside the system (point B), into the discharge area inside the system (point C), and out of the system in the discharge area (point D). This systematic movement of water is caused by a driving force, the hydraulic gradient established by the water table topography in conjunction with the pull of gravity on water from areas of higher to areas of lower hydraulic potentials (Hubbert, 1940; Tóth, 1962, 1963).

Like other forms of flow, regional groundwater flow has an input and an output. The input in this case, is provided by infiltration from precipitation which in most cases is areally extensive, but occasionally is temporally and spatially concentrated. The output occurs as discharge to fluvial systems and evapotranspiration all over the region (Figure 3).

Several points may be noted from Figure 3, and a consideration of flow distribution in the system: 1. the groundwater system consists of subsurface and surface (precipitation and evapotranspiration) subsystems; 2. the direction of movement of precipitation at the water table is similar to the direction in which water moves into the system in the recharge area, but is opposite to the direction of the exit of water in the discharge area; 3. water at point B in the recharge area will get to point C in the discharge area, eventually. Consequently, some conclusions can be drawn intuitively: First, under the same conditions of rainfall, it would appear as if more water will infiltrate into the system through the recharge area (point A) than through the discharge area (point D). In fact, a disturbance in the form of development may be necessary to reverse the natural hydraulic gradient, in order to induce infiltration into the system in the discharge area. In the recharge area, such a disturbance will only enhance the rate of infiltration into the system. Secondly, due to the continuous movement of groundwater from the recharge area to the discharge area, a perennial surplus of moisture as compared to the amount of ground moisture expected from the balance of precipitation and evapotranspiration could exist in the discharge area. In the recharge area, a deficit equivalent to this surplus should exist. In other words, there could be more water in the discharge area than in the recharge area of a basin, under natural conditions.

The second conclusion above has for long been recognized by various scientists, who have applied it in the solution of water resources and related problems. Meinzer (1923, 1927), studied the hydraulic discharge of water, using plants and other indicators of groundwater surplus. That saline soils are indicative of upward rising groundwater was discussed by Jennings et al. (1934), and Benz et al. (1961) in Utah and North Dakota, respectively. Meyboom (1963, 1966), cited several surface manifestations of groundwater discharge in different parts of the Canadian prairies. It was Tóth (1966a) who developed the relationship between groundwater

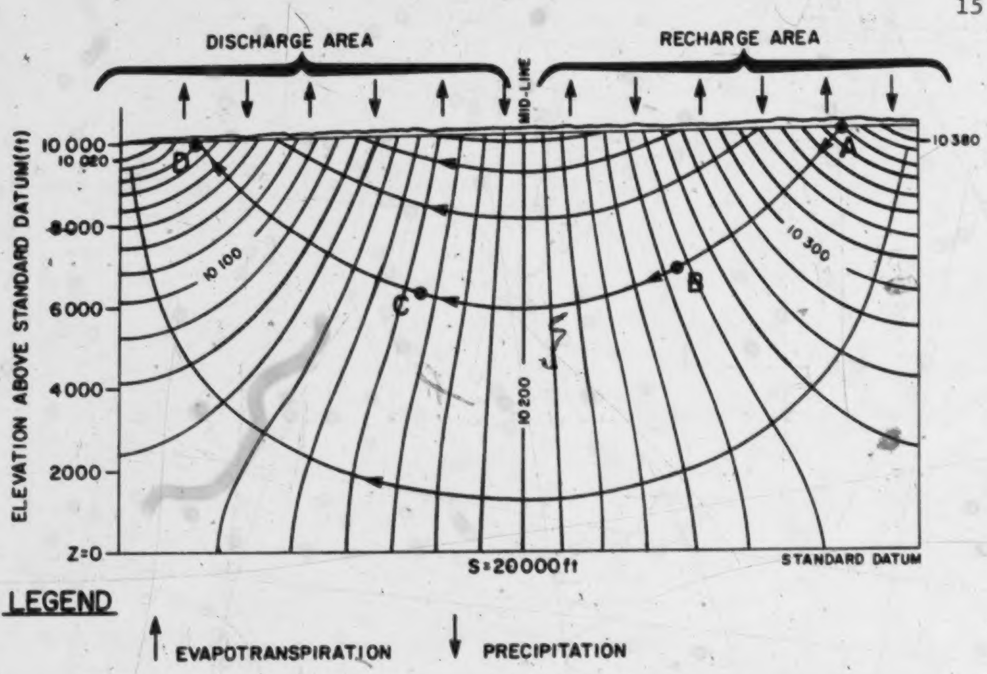

DISCHARGE AREA
RECHARGE AREA
MID-LINE
ELEVATION ABOVE STANDARD DATUM(ft)
10 000
10 020
8000
6000
4000
2000
Z=0
10 380
10 300
10 100
10 200
A
B
C
D
S=20000ft
STANDARD DATUM
LEGEND
EVAPOTRANSPIRATION
PRECIPITATION

Figure 3. Discharge, midline and recharge areas of a groundwater flow system (after Tóth, 1962).

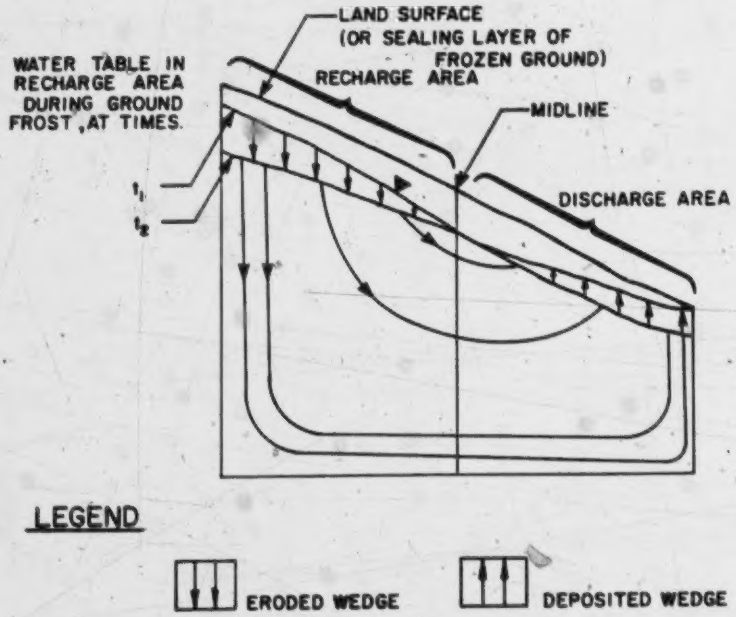

LAND SURFACE (OR SEALING LAYER OF FROZEN GROUND)
WATER TABLE IN RECHARGE AREA DURING GROUND FROST, AT TIMES.
RECHARGE AREA
MIDLINE
DISCHARGE AREA
$t_1$
$t_2$
LEGEND
ERODED WEDGE
DEPOSITED WEDGE

Figure 4. Schematic distribution of water between recharge and discharge areas below a frozen layer (after Tóth, 1982).

flow and surficial indicators into a groundwater mapping tool in the light of his theory (Tóth, 1963).

As an agent of transportation and deposition, the movement of groundwater from recharge to discharge areas has also been employed in important studies. Tóth (1980), in his proposition of the theory of petroleum migration, states that hydrocarbon deposits are moved toward discharge foci, where they are preferentially associated with ascending limbs and stagnation zones of flow systems. In their analysis of the genesis of ore deposits, Garven and Freeze (1982) concluded that regional flow systems provide a feasible mechanism for leaching large volumes of source beds, focusing the flow into aquifers and transporting metal species to depositional sites.

Unfortunately, these studies seem to have skipped one basic step, that of a systematic quantification of groundwater across the regional system. The argument here is that, as water transports hydrocarbons and other ore-forming minerals along the flow paths, water also transports itself in the process. Unlike these minerals, however, no specific depositional mechanisms are associated with water. Nevertheless, an uneven quantitative distribution of water between the recharge and discharge area is conceivable. Figure 4 is a modified schematic distribution of water between the recharge and discharge areas of a basin (Tóth, 1982), below a frozen upper layer. One could conceive of the depleted wedge above the recharge area as that part of an eroded mineral deposit, which is then deposited in the wedge above the discharge area. This statement is not academic. In the field, the depleted and deposited wedges are marked by drawdown and rise in water levels in observation wells, respectively. A systematic distribution of groundwater resources across a basin is anticipated.

Along these lines, Tóth (1966b) believes that an established picture of groundwater motion can be used for locating areas with the most favourable conditions of hydraulic potential. Thus, Tóth located his first test hole in the discharge area during his exploration for groundwater near Olds in the Plains Regions of Alberta. However, the lack of a proper documentation of this phenomenon has often led to silence over it, in cases where it was observed (Stephens, 1985). Stephens studied the cumulative increase in soil water storage at the crest, slope and flank of an active sand dune in response to a precipitation storm. It was observed that 50, 100 and 200 percent of the precipitation accumulated in the soil at the crest, slope and flank, respectively. He concluded that the 200 percent at the flank was contibuted by lateral flow from adjacent areas.

The systematic distribution of groundwater, and the variation in the infiltration of precipitation across the basin have, thus far, been considered under natural steady-state conditions. The development of the basin might result in changes in the postulated patterns of groundwater distribution. Furthermore, in areas as large as the Plains Regions of Alberta, variations in aspects of the hydrogeologic environment, such as topography, geology, and climate could be expected to complicate any simple pattern of groundwater distribution that may exist.

In the southern part of the Plains Regions where any quantity of water appears to be important, it would be necessary to plan the exploitation of groundwater in advance. If the basin is to remain as the unit of groundwater studies, as is the case in some other parts of the world (Gutentag, 1985; Lindholm, 1985; Harrill, 1985; Wilkins, 1985; Taylor, 1985), a predetermined systematic approach to the optimal exploitation of regional groundwater will be desirable. It will be the author's task in the chapters that follow to develop this approach for a variety of the hydrogeologic environments which occur in the Plains Regions of Alberta.

## 1.4 Objectives

This work is focussed upon utilizing numerical simulation models of the Plains Regions environment in Alberta to show how regional groundwater flow may be incorporated in management models. Flow distribution within the system is of major concern. The first objective is to show that Plains Regions environments can be characterized. Work on this problem is essentially the application of the concept of hydrogeologic environment. The second goal is to determine if flow conditions can be treated and quantified in a logical manner. A hydrological (baseflow) analysis, followed by theoretical steady-state models, and then the transient development of unit basins of the region serve as tools for achieving this goal.

The third objective is to establish quantitatively if any relationship exists between the occurrence of water and basin geometrical configuration (discharge-recharge areas). This is done by comparing results of the second objective with the discharge-recharge regimes of the system. Basin stability is also considered at this stage.

The final objective is to arrange any relationships in the form of guidelines for regional evaluation and optimal management practices in the Plains Regions. This objective is to be established in the form of a sensitivity analysis.

This analysis is made particularly attractive because of the wide variety of environments encountered in the Plains Regions of Alberta. The methodologies and results of this study should have widespread transfer value, since the hydrogeologic parameters of the environments in this region correspond to those of most glaciated and some unglaciated parts of the world.

### 1.5 Scope

This work is on the development and management of regional groundwater resources in the Plains Regions of Alberta. The emphasis has been on a good understanding of conditions of regional groundwater flow in this region, and its inclusion in a systematic development and management model. In order to approach this problem, a model must be capable of handling the complex water system in this region. To this end, some existing numerical models were selected as the principal tools in developing the required understanding.

The chapters which follow are on the development and discussion of the methodology of the inclusion of the consequences of regional groundwater flow in management models.

In chapter 2, a hydrogeological and hydrological characterization of the Plains Regions environment is given.

A theoretical steady state characterization of the environment is done in chapter 3.

In chapter 4, a transient analysis is done during which type basins are developed systematically by the placement of one well at a time to test the results of chapters 2 and 3. The effects of changes in environmental parameters and hence their role in basin planning and management are considered also.

Chapter 5 is focussed upon the application of the developed models in the management of groundwater in Ross Creek Basin.

Chapter 6 is a summary of the major findings.

# 2. CHARACTERIZATION OF PLAINS REGIONS ENVIRONMENT

## 2.1 Hydrological characterization

One problem that is usually encountered in studies involving areas as large as the Plains Regions of Alberta (Figure 5) is that the area is too extensive to be treated as one unit of study. A description of the parameters based on the consideration of every existing report, will include numerous reports that cannot be handled practically. It becomes necessary that a standard procedure be set with which the study area can be characterized. Such a procedure must be relevant and adequate to the objectives of this study.

For the reasons outlined above, baseflow analysis was selected as the basic procedure to aid in the characterization of the Plains Regions environments. Baseflow analysis is relevant to the present study because baseflow is an element of regional groundwater flow. The adequacy of baseflow analysis to this study lies in the fact that baseflow is controlled by the climate, topography and geology of the setting. Thus, analysis of baseflow integrates the effects of these three important components of the hydrogeologic environment. In this regard, the analysis of baseflow has been employed as a tool in several studies associated with regional groundwater resources (Searcy, 1959; Farvolden, 1963; Amus'ya and Ratner, 1964; Sharov, 1965; Popov, 1965).

In addition to its proposed role as a characterizing technique, it was also hoped that the analysis of baseflow would provide a basic understanding of the nature of regional groundwater discharge to streams in the Plains Regions. Moreover, the data required for the analysis of baseflow are readily available from the many stream gauging stations in the Plains Regions. The streamflow records were obtained from files of Alberta Environment, Edmonton.

### 2.1.1 Baseflow analysis

Baseflow is the portion of streamflow that is contributed by groundwater discharge. Baseflow often forms an insignificant percentage of streamflow. However, it is important in water balance studies because it comprises a part of that portion of precipitation in an area which takes part in the complete hydrologic cycle.

Baseflow may be separated from the surface runoff component of stream hydrographs by the several graphical methods discussed by Meyboom (1961a), Chernaya (1964), and Linsley

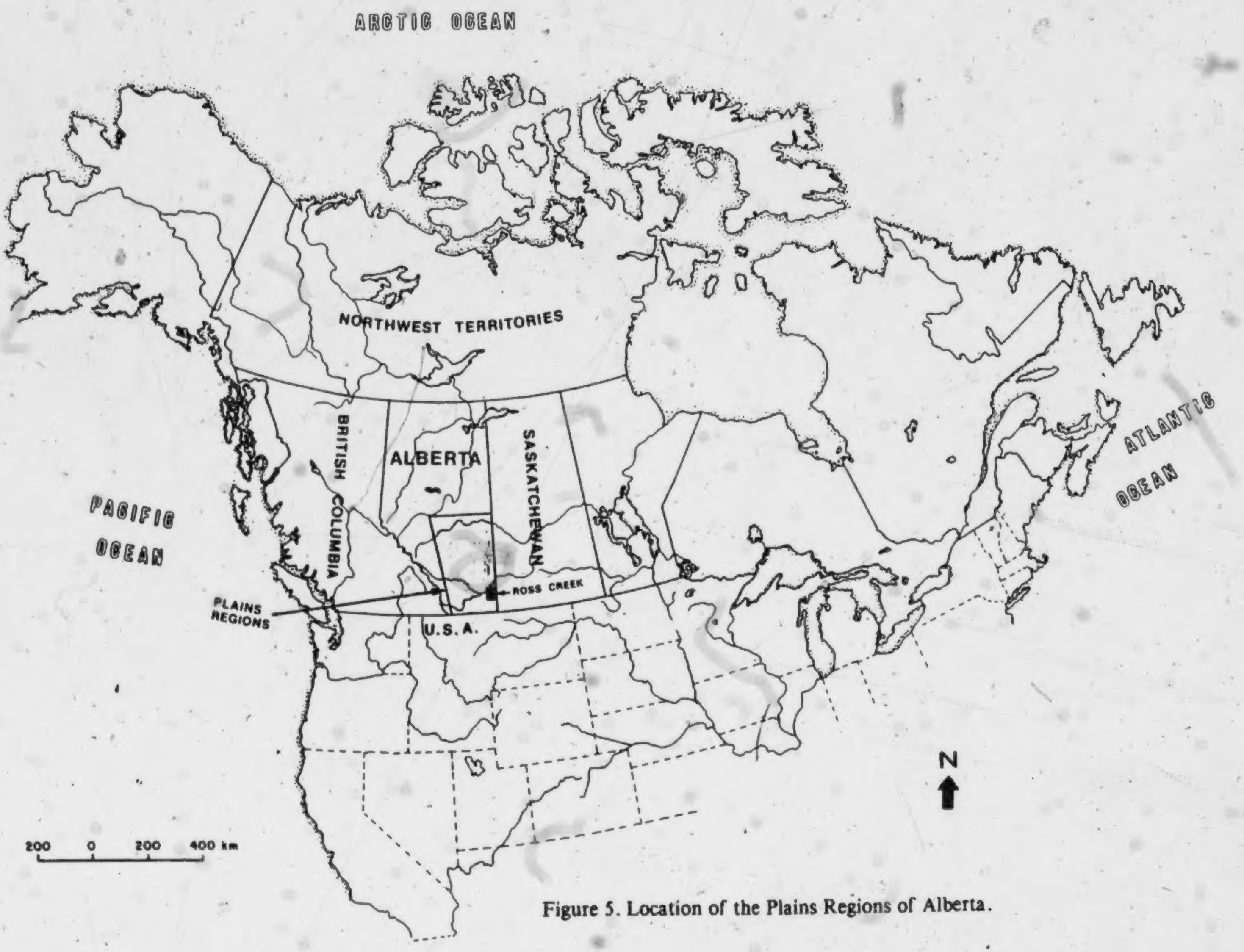

ARCTIC OCEAN
NORTHWEST TERRITORIES
BRITISH COLUMBIA
ALBERTA
SASKATCHEWAN
PACIFIC OCEAN
ATLANTIC OCEAN
PLAINS REGIONS
ROSS CREEK
U.S.A.
N
200 0 200 400 km

Figure 5. Location of the Plains Regions of Alberta.

et al. (1975), to mention three. These methods are not only difficult to use, but are also hard to employ consistently. For this reason, only the three methods to be discussed later were used in the present study. As the purpose of the baseflow determination was comparative, emphasis was laid on consistency, so long as reasonable precision was attained.

The amount of discharge through a basin is dependent on several factors, such as the infiltration capacity of the surficial deposits, the amount, nature, and distribution of the precipitation, and certain physical properties of the basin. Spence (1971) analyzed the influence of up to 39 measures of climatic and physical geographic patterns on streamflow in the Plains of Canada. In this study, it is assumed that these factors are relatively uniform across the study area, and that the only variable is the area of the basins. Although this assumption is subject to criticism, it is believed that any relationship obtained in the study would be valid because of the random selection of the basins considered. Moreover, Spence (1971) concluded that the contributing drainage area is the single most important variable controlling streamflow.

Records of all streamflow gauging stations in southern Alberta from the North Saskatchewan River to the Alberta-Montana (U.S.A.) boundary were assembled and examined. In choosing the stations, and hence the basins to be used in the study, the following factors were considered: 1. records should cover the same period of time; 2. gauging should be continuous rather than seasonal; 3. the range in basin areas should be acceptable for most Plains Regions basins in Alberta; 4. the basins should be selected in order to minimize the effect of regulation on streamflow; and 5. the basins should lie in the Plains Regions as much as possible. The combined consideration of these factors made the basin selection a formidable task, and perhaps, almost impossible. Consequently, some of the requirements were relaxed, and thirty-two basins were selected a shown in Figure 6.

#### 2.1.1.1 Empirical baseflow curves

To the author's knowledge, there appears to be no completely satisfactory method for the separation of the base- and surface-runoff components of the often intermittent streamflow of most Plains Regions streams. The three methods used herein were selected to facilitate the large number of computations. These methods are approximate, namely: minimum monthly discharge, "cutoff along the low water", and baseflow recession.

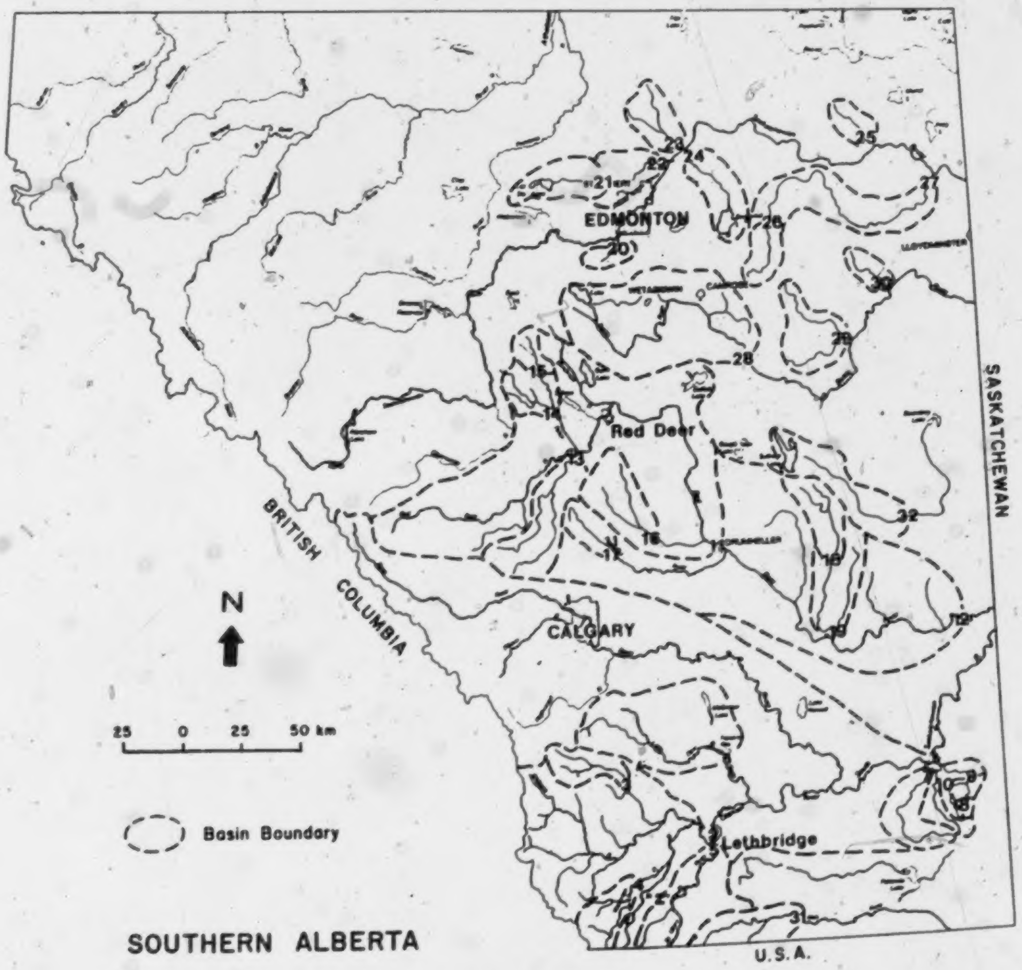

SASKATCHEWAN
BRITISH COLUMBIA
U.S.A.
EDMONTON
Red Deer
CALGARY
Lethbridge
21
22
23
24
25
26
27
20
30
29
28
15
14
13
17
16
18
19
32
12
8
10
9
31
N
25
0
25
50 km
Basin Boundary
SOUTHERN ALBERTA

Figure 6. Selected basins for baseflow analysis.

#### 2.1.1.1.1 Minimum monthly discharge method

It has long been recognized that the minimum temporal discharge in a stream is an indication of groundwater discharge into the stream (Houk, 1921; Rasmussen and Andreasen, 1959). A monthly minimum-flow time consideration has been introduced and used in various forms to estimate baseflow in the Russian literature since the 1960's (Chernaya, 1964). In the Canadian prairies, Freeze (1969a) used minimum monthly stream discharges to estimate annual baseflow for several streams. This method is simple and easily applicable. The flashy intermittent nature of most of the streams in the Plains Regions is no hinderance to the applicability of this method, but baseflow values obtained in this manner must be used with caution since Laycock (1957) pointed out that melted snow sustains streamflow in some rivers during extended periods of dry weather. However, the method was used to estimate baseflow in all the thirty-two selected basins for all years of record between 1965 and 1982, because such values of baseflow can be correlated with average annual groundwater recharge (Freeze, 1969a). For each basin, monthly volumes (Table 1) were estimated by multiplying the recorded minimum daily discharges by the number of days in the month. An average minimum annual discharge (baseflow) was then calculated after the conversion of monthly volumes to annual volumes. Following this, the baseflow was divided by the area of the basin to obtain the baseflow rate per unit area (specific baseflow rate, q), as posted in Table 2. Next, the specific baseflow rates were plotted against the areas (A) of the basins (Figure 7). Finally, a smooth line of best fit was drawn through the plotted points.

#### 2.1.1.1.2 Cutoff along the low water method

This is a simple graphical stream hydrograph separation method which employs arithmetic plots of stream hydrographs, and a line drawn along the low water level (Chernaya, 1964). In the present study, the streamflow records of the period from 1978 to 1982 were analyzed by this method. This period was chosen because some of the gauging stations had no records prior to 1978. In the instances where more records were available, an examination of the data indicated that the records between 1978 to 1982 were in agreement with data of typical years as conceptualized by Kunkle (1962), and Farvolden (1963).

For each basin, annual arithmetic plots of stream hydrographs were generated as exemplified by Figure 8. A line drawn through the minimum low water level, or "the low

Table 1. Esimation of baseflow from minimum monthly discharge.

| Year | Minimum monthly volume (x $10^{-6}$ $m^3$) | | | | | | | | | | | | Total minimum annual volume (x $10^{-6}$ $m^3$/yr) | Minimum annual disch. (baseflow) (x $10^{3}$ $m^3$/s) |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | Jan | Feb | Mar | Apr | May | Jun | Jul | Aug | Sep | Oct | Nov | Dec | | |
| 1976 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.02 | 0.02 | 0.02 | 0.02 | 0.0 | 0.0 | 0.08 | 2.63 |
| 1977 | 0.0 | 0.0 | 0.0 | 0.04 | 0.22 | 0.0 | 0.01 | 0.01 | 0.01 | 0.02 | 0.0 | 0.0 | 0.31 | 10.33 |
| 1978 | 0.0 | 0.0 | 0.0 | 0.13 | 0.35 | 0.09 | 0.0 | 0.0 | 0.01 | 0.02 | 0.0 | 0.0 | 0.60 | 18.85 |
| 1979 | 0.0 | 0.0 | 0.0 | 0.0 | 0.63 | 0.56 | 0.86 | 0.12 | 0.01 | 0.01 | 0.0 | 0.0 | 2.18 | 69.21 |
| 1980 | 0.0 | 0.0 | 0.0 | 0.0 | 0.12 | 0.02 | 0.0 | 0.03 | 0.02 | 0.01 | 0.0 | 0.0 | 0.20 | 6.82 |
| 1981 | 0.0 | 0.43 | 0.04 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.47 | 14.79 |
| 1982 | 0.0 | 0.0 | 0.0 | 0.0 | 0.10 | 0.0 | 0.0 | 0.0 | 0.0 | 0.01 | 0.0 | 0.0 | 0.11 | 3.45 |
| Average minimum annual discharge (baseflow) | | | | | | | | | | | | | = | 18.01 |

* Station 05ED002, Atimoswe Creek near Elk Point

Table 2. Empirical specific baseflow rates.

| Basin number | Station number | Area of basin (km²) | Specific baseflow rate (x $10^3$ m³/s/km²) | | |
|---|---|---|---|---|---|
| | | | Minimum monthly discharge method | Cutoff along low water method | Baseflow recession method |
| 1 | 05AJ001 | 56 400 | 1.740 | 0.556 | 0.37 |
| 2 | 05AD007 | 17 000 | 2.312 | 0.307 | 0.53 |
| 3 | 05AB021 | 1 160 | 1.085 | 0.075 | 0.10 |
| 4** | 05AD028 | 1 600 | 5.269 | 0.800 | 0.63 |
| 5 | 05AE006 | 3 530 | 1.243 | 0.592 | 0.36 |
| 6 | 05AC012 | 5 330 | 0.059 | -- | 0.0* |
| 7 | 05AH005 | 3 270 | 0.067 | 0.0 | 0.0* |
| 8 | 05AH046 | | 0.0 | 0.0 | 0.0* |
| 9 | 05AH003 | 624 | 0.037 | 0.0 | 0.0* |
| 10 | 05AH037 | 215 | 0.031 | 0.0 | 0.0* |
| 11 | 05CE001 | 24 800 | 1.107 | 0.300 | 0.32 |
| 12 | 05CK004 | 44 700 | 0.719 | 0.110 | 0.0 |
| 13 | 05CB001 | 2 560 | 0.706 | 0.039 | 0.16 |
| 14 | 05CC007 | 1 910 | 0.502 | 0.430 | 0.25 |
| 15 | 05CC010 | 54 | 0.523 | 0.0 | 0.08* |
| 16 | 05CE002 | 2 430 | 0.049 | 0.0 | 0.0* |
| 17 | 05CE006 | 753 | 0.066 | 0.0 | 0.0* |
| 18 | 05CH008 | 1 550 | 0.014 | 0.0 | 0.0* |
| 19 | 05CH007 | 3 720 | -- | 0.0 | 0.0* |
| 20 | 05DF006 | 335 | 0.040 | 0.0 | 0.0* |
| 21 | 05EA005 | 1 910 | 0.515 | 0.274 | 0.27 |
| 22 | 05EA001 | 3 350 | 0.457 | 0.043 | 0.07* |
| 23 | 05EC005 | 1 550 | 0.435 | 0.046 | 0.06* |
| 24 | 05EB015 | 2 890 | 0.003 | 0.0 | 0.0* |
| 25 | 05ED002 | 364 | 0.049 | 0.0 | 0.0* |
| 26 | 05EE003 | 1 590 | 0.034 | 0.0 | 0.0* |
| 27 | 05EE007 | 7 270 | 0.030 | 0.0 | 0.0* |
| 28 | 05FC001 | 7 680 | 0.203 | 0.047 | 0.02* |
| 29 | 05FB002 | 3 500 | 0.028 | 0.0 | 0.0* |
| 30 | 05FE002 | 709 | 0.069 | 0.0 | 0.0* |
| 31 | 11AA005 | 2 720 | 2.194 | 0.083 | 0.05 |
| 32 | 05GA008 | 3 000 | 0.003 | 0.0 | 0.0* |

* read from calibration curve with minimum monthly flow method.

** omitted from graph, lies wholly in the mountain regions.

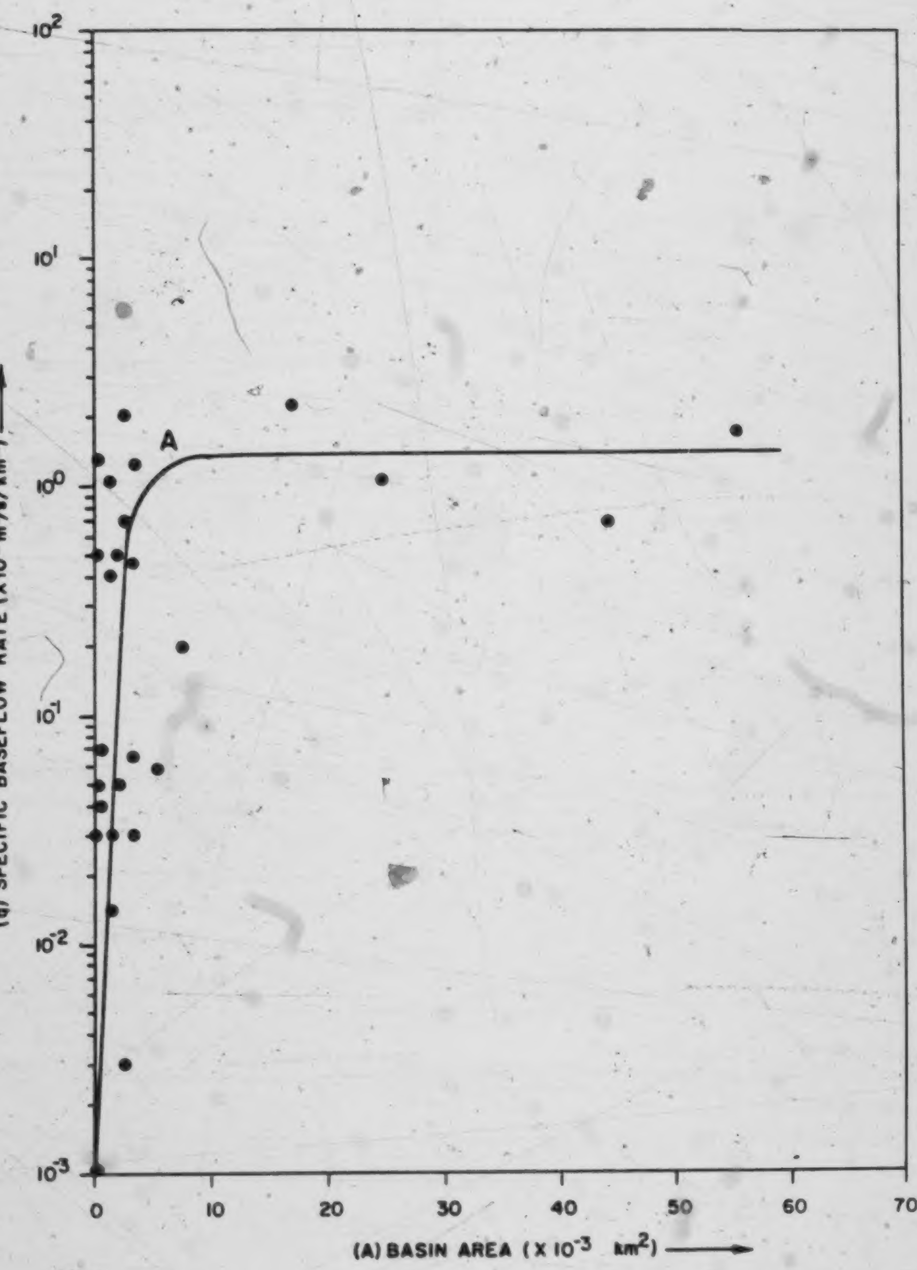


Figure 7. Empirical q(A) curve by minimum monthly discharge method.

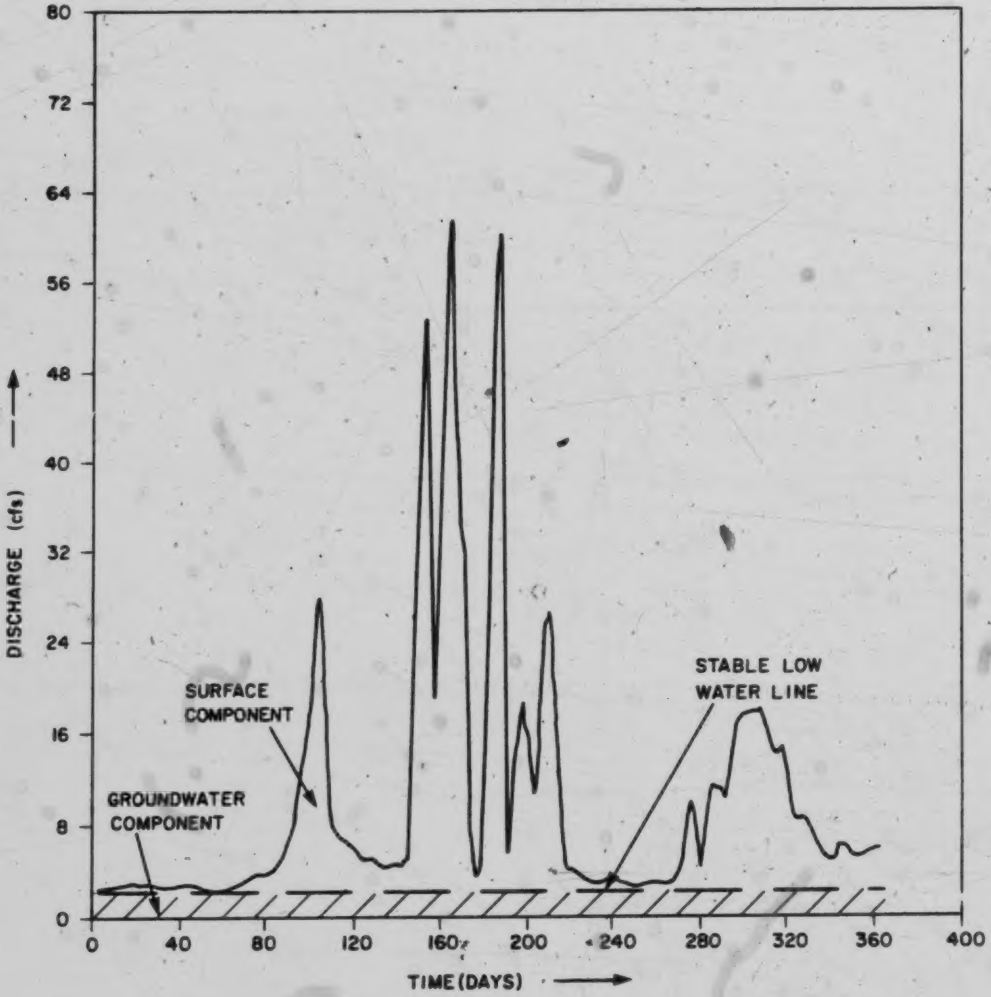


Figure 8. Stream hydrograph for cutoff along low-water method.

water" separates the baseflow from the surface runoff component of the hydrograph (Figure 8). To estimate the baseflow, a transparent graph sheet containing squares of known area was superimposed on the hydrograph. The squares lying in the baseflow field were then counted and converted to units of baseflow. The estimated baseflow was divided by the area of the basin to produce the specific baseflow rate (Table 2). A plot of specific baseflow rate versus basin area was constructed using only the non-zero values, and a best fitting line was drawn through the points (Figure 9).

#### 2.1.1.1.3 Baseflow recession method

In the presence of the hydraulic gradient created by the water table, groundwater flows toward the channel in a region. During extended periods of dry weather, continuous discharge of groundwater into the channel lowers the water table over the uplands area, and decreases the hydraulic gradient and water movement according to Darcy's Law.

The depletion of groundwater discharge is reflected in the falling limb or recession of the stream hydrograph. The curve may be expressed by the depletion equation (Linsley et al., 1975):

$$q_1 = q_0 K_r \qquad (11)$$

where: $q_0$ = the flow at any time (acre-ft/day);
$q_1$ = the flow one time unit later (acre-ft/day);
$K_r$ = the recession constant, which is less than unity.

In a more general form, the recession could be written as (Barnes, 1939):

$$q_t = q_0 K_r^t \qquad (12)$$

where: $q_t$ = the flow t time units after $q_0$ (acre-ft/day);
t = time interval (days).

Buttler (1957) wrote the recession equation:

$$Q = \frac{K_1}{10^{t/K_2}} \qquad (13)$$

where: Q = the flow at any time (acre-ft/day);

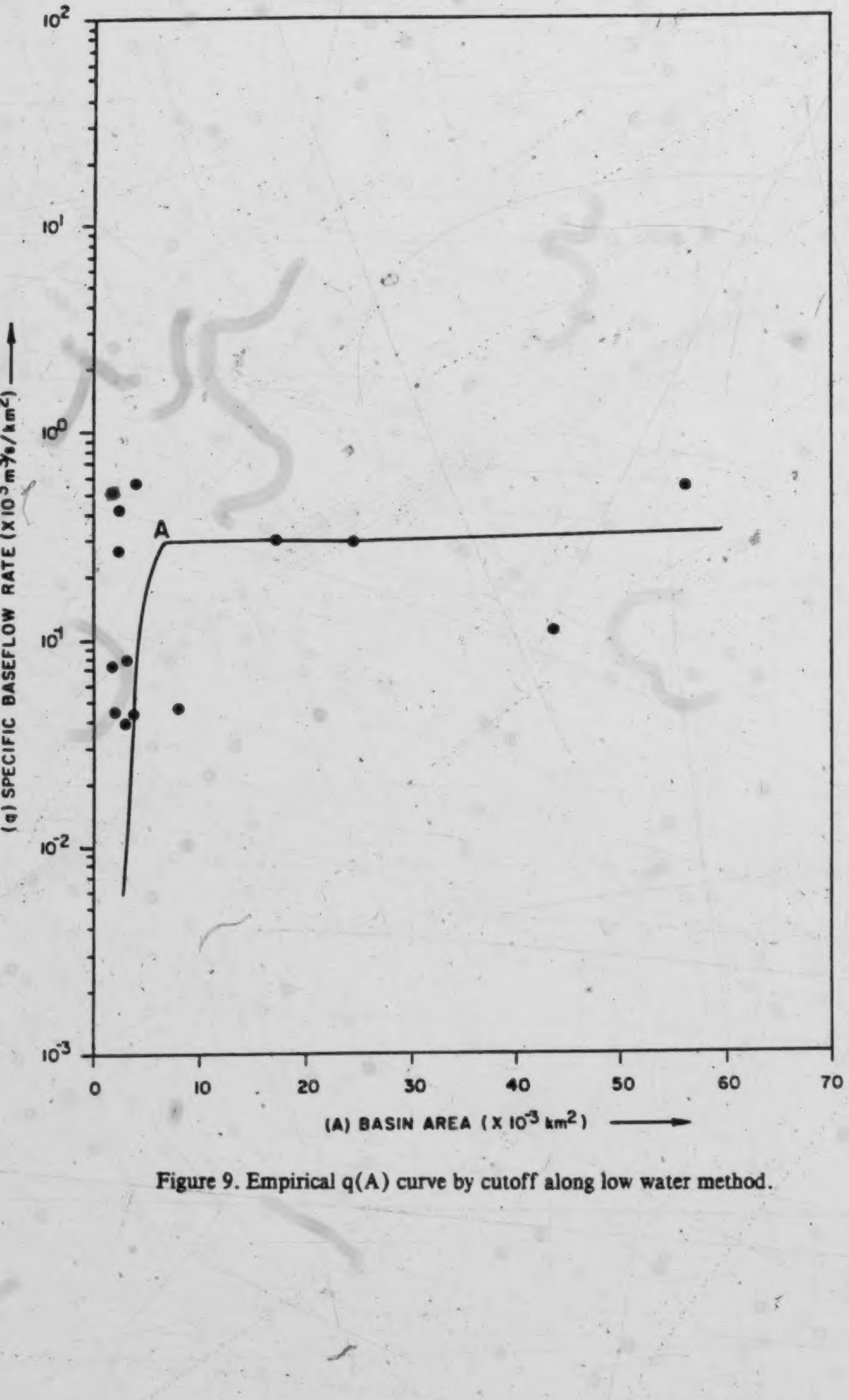


Figure 9. Empirical q(A) curve by cutoff along low water method.

$K_1$ = Q at t = 0 (acre-ft/day);

$K_2$ = t where Q = 0.1 $K_1$ (days).

and showed that the actual volume of discharge (Qa) in a given time interval could be obtained by integrating (13) as:

$$Qa = \int_{t_1}^{t_2} Qdt = \left[\frac{-K_1K_2/2.3}{10^{t/K_2}}\right]_{t_1}^{t_2} \qquad (14)$$

where: Qa = volume of discharge (acre-ft).

Considering $t_1=0$ at the beginning, and $t_2$=time at the end of a recession period, (14) reduces to:

$$Qa = \left[\frac{K_1K_2}{2.3}\right] - \left[\frac{K_1K_2/2.3}{10^{t_2/K_2}}\right] \qquad (15)$$

The first term on the right hand side of (15) corresponds to the "total potential groundwater discharge", and the second term to the "remaining potential groundwater discharge" of Meyboom (1961a).

Equation (12) shows that if the logarithm of the discharge is plotted against time, a linear relationship between the two variables may be obtained, with the best straight line representing groundwater recession. To calculate the amount of natural groundwater discharge for each of the basins, semilogarithmic plots of the streamflow data were constructed (Figure 10). For the same reasons outlined in the preceding section, only the records for the period from 1978 to 1982 were considered. For each basin, and for each year, (15) was used to determine the actual groundwater discharge. The start and end of the recession curve, and the values of $K_1$ and $K_2$ were obtained by analyzing the stream hydrograph in the same manner as Meyboom (1961a).

It must be noted that this method of stream hydrograph separation, while useful in dealing with permanent streams, is not suited for some of the streams considered in this study because of their intermittent nature. Consequently, the method was employed in the separation of the hydrographs of the eleven permanent streams only. For the remaining twenty-one intermittent streams, equivalent values were calculated from a calibration curve relating the two

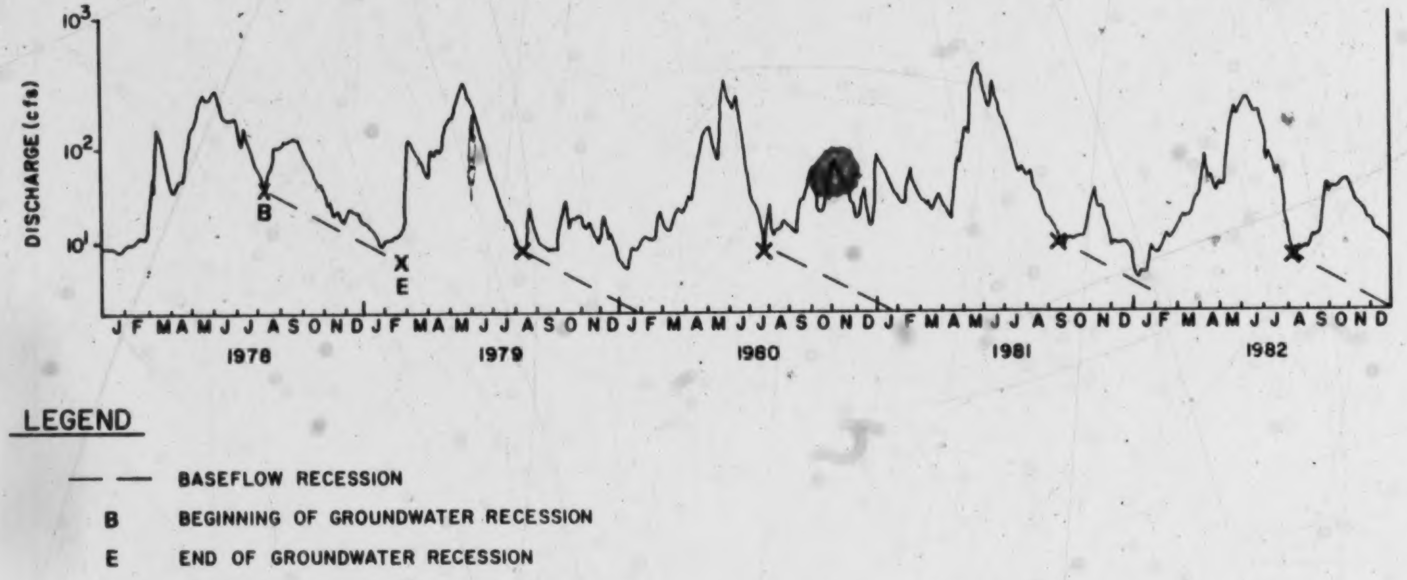


Figure 10•Stream hydrograph for separation by baseflow recession method.

sets of baseflow values estimated for the permanent streams by this method and the minimum monthly discharge method. To obtain the calibration curve, the baseflow values for the eleven permanent streams, obtained by the two methods were plotted (Figure 11). Using the method of least squares, an equation for the line of best fit was determined, and the line was drawn through the points. With the aid of the calibration curve, crude baseflow estimates were then made for the intermittent streams.

In all the instances, the baseflow values were divided by the corresponding areas of the basins to obtain the specific baseflow rates which are displayed on Table 2, and the non-zero values plotted (Figure 12). As with the other two methods, a line of best fit has been drawn through the points.

### 2.1.2 Results and implications of baseflow analysis

#### 2.1.2.1 Estimated baseflow values

Table 2 reveals a certain level of consistency in the values of baseflow estimated by the three different methods. Thus, the values can be considered as realistic approximations to true groundwater discharge to streams in the Plains Regions of Alberta. However, the minimum monthly discharge method appears to overestimate baseflow, relative to the methods of "cutoff along low water" and baseflow recession. This difference is conceivable because the method of minimum monthly discharge produces values which may or may not represent actual groundwater discharge to streams (Freeze, 1969a).

#### 2.1.2.2 Baseflow regularity

The q(A) curves which were constructed through the plotted points (Figures 7, 9, and 12), display a regularity. This regularity is such that, with increasing area of basin at initial low values of basin area, specific baseflow rate increases rapidly. At and above a certain size of basin area (point A), the rate of increase of specific baseflow rate decreases with further increase in the area, thus specific baseflow rate tends toward a constant value above point A. This property of the specific baseflow rate will be referred to as "baseflow regularity" henceforth in this thesis.

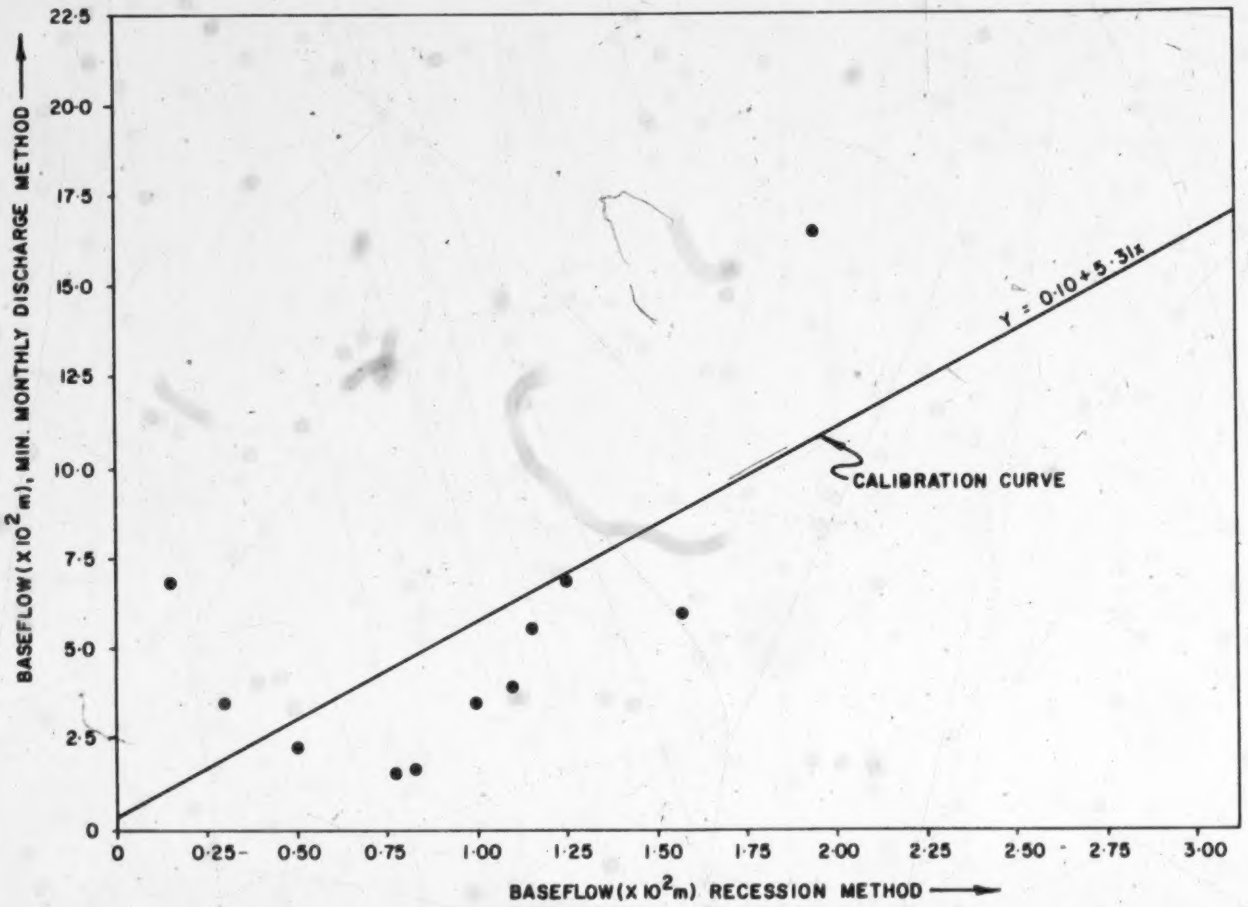


Figure 11. Baseflow calibration curve using permanent stream values.

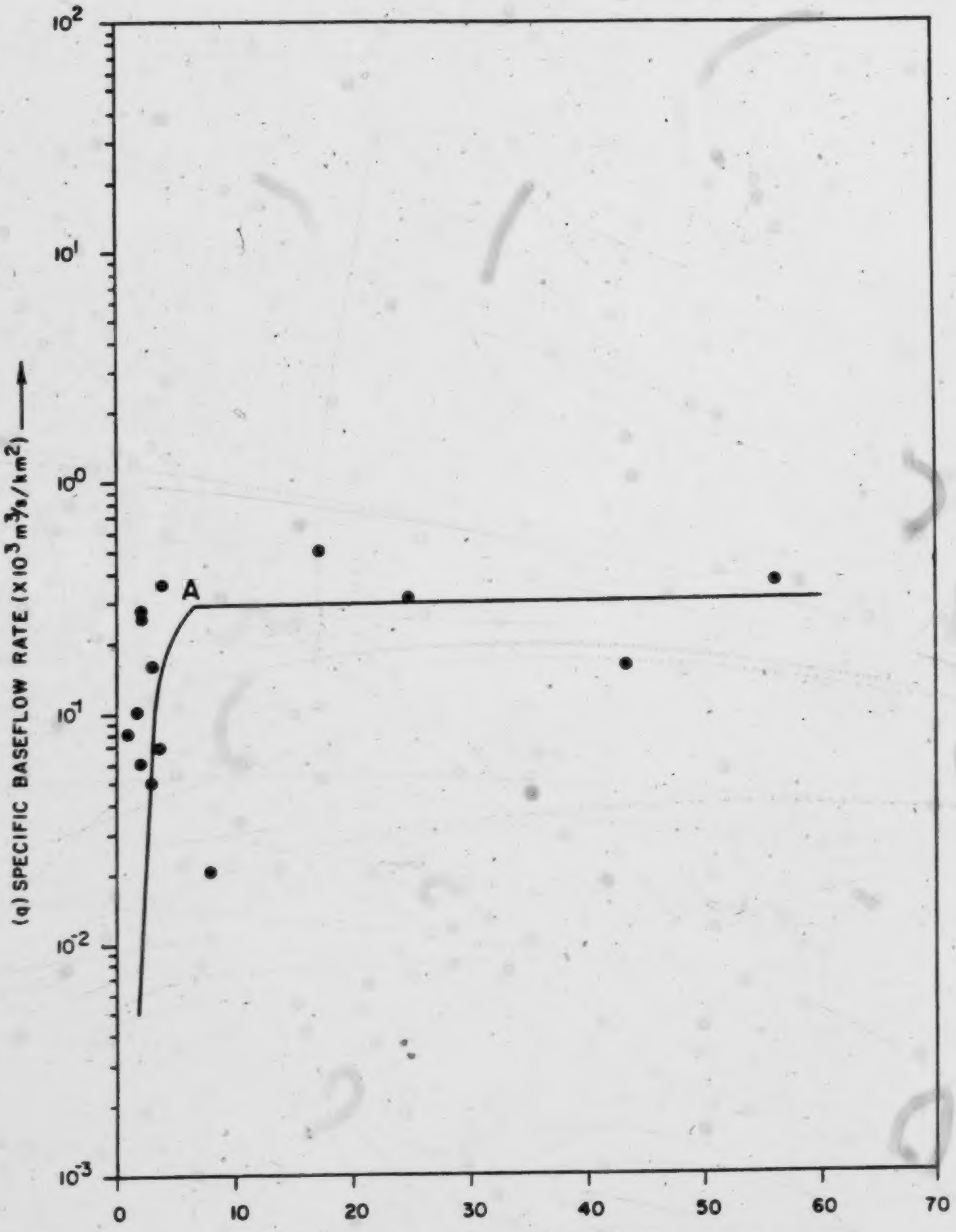


Figure 12. Empirical q(A) curve by baseflow recession method.

The observation of regularities in groundwater discharge is not completely new in hydrological studies. Popov (1965) reported the existence of such regularities in his study of the formation of underground flow into rivers on the territory of the USSR (Figure 13). He did not, however, explain these regularities in the context of regional groundwater flow.

The baseflow regularity obtained in the present analysis indicates several aspects relating to regional groundwater flow in the Plains Regions, as portrayed by regional groundwater discharge to streams: 1. groundwater discharge occurs in a systematic manner; 2. all other factors being equal, the size of the basin appears to control groundwater discharge up to a certain magnitude; 3. above a certain size of basin (approximately 6 000 km$^2$, corresponding to the inflection point A in Figures 7, 9 and 12), conditions of groundwater discharge become relatively constant, therefore, basins of and above 6 000 km$^2$ may be considered as "type units" or "type environments" for regional groundwater studies in this region.

The scatter of the points about the q(A) curves in Figures 7, 9 and 12, is noteworthy, and it sheds exceptions on the general rules in the preceding paragraph. This scatter is thought to have resulted from the lack of a detailed consideration of all the factors which affect the discharge of groundwater to streams, namely: the artificial regulation of streamflow, effective drainage area, and other physical and chemical properties of the hydrogeologic environment. A detailed consideration of all the factors is beyond the scope of the present study, but those that are considered to be most important in the Plains Regions will be evaluated in a later section.

### 2.1.2.3 Hypothesis for the distribution of regional groundwater discharge

On the basis of the baseflow regularity, a systematic groundwater discharge distribution pattern is proposed: in the Plains Regions of Alberta, and at the large scale at which the baseflow analysis was carried out, as one moves from high order streams through lower streams to the stream of lowest order, one transects sub-basins of increasing area from high to low elevation (Figure 14). These sub-basins are often bounded by small regularly occurring topographic divides with their streams occupying their lower central portions (Figure 15, points G, F, C, D). The water table which often coincides with the surface topography (Jones, 1960; Farvolden, 1961; Meyboom, 1961b), generates flow systems of different order (local, intermediate, regional) as indicated in Figure 15. Water recharged into the system at point A, is

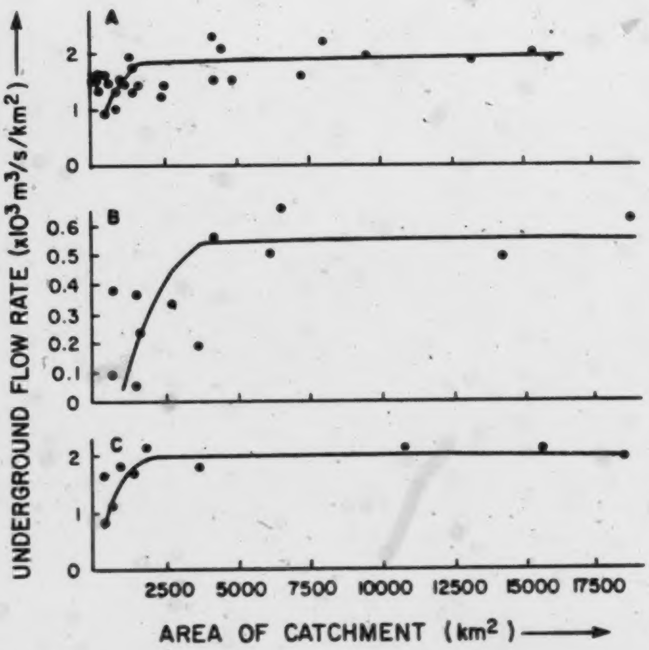


Figure 13. Rates of underground flow into rivers versus area of catchment for different hydrogeologic regions in the USSR (after Popov, 1965)

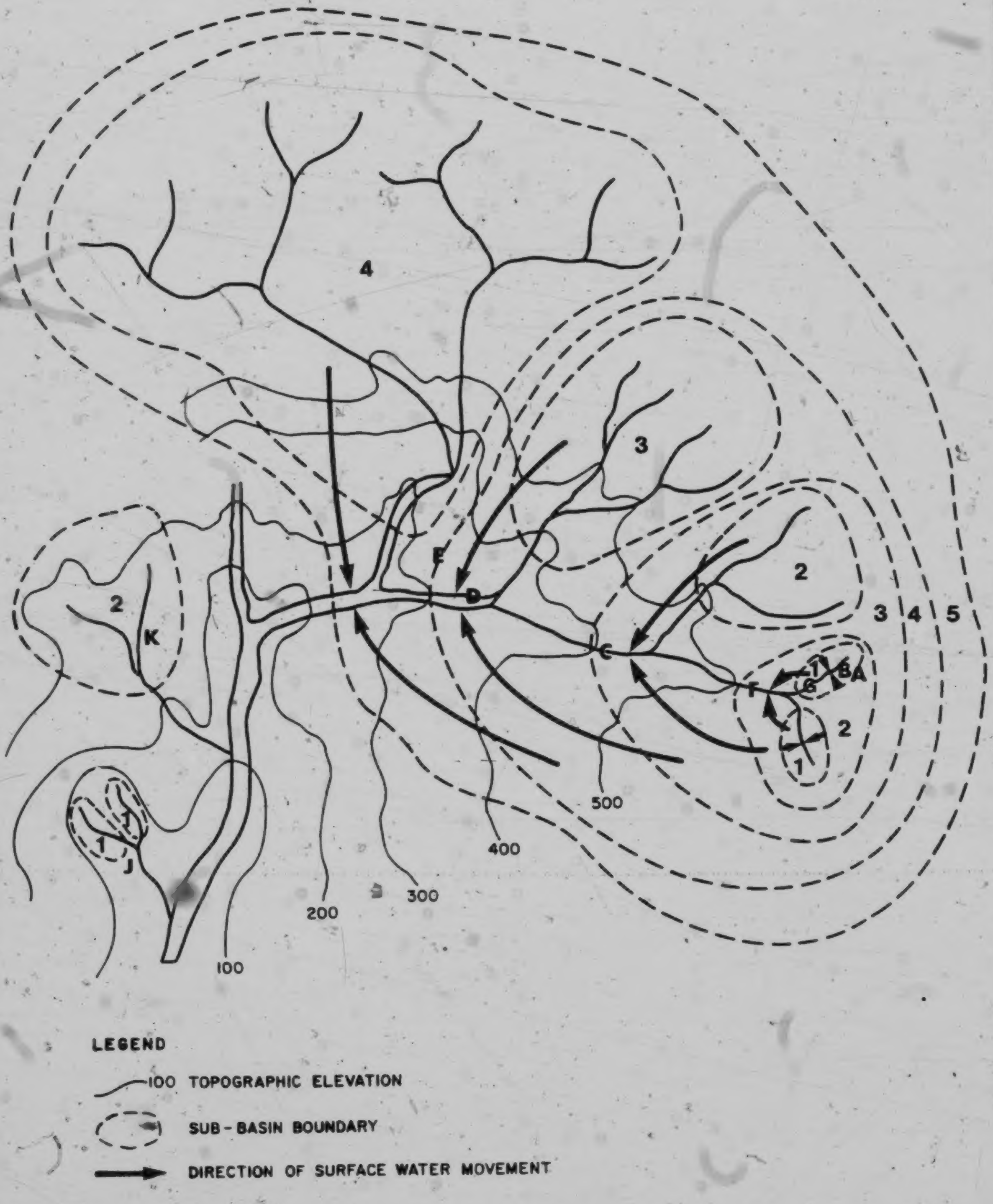

4
3
2
2
3 4 5
E
D
C
F
G
B
A
1
2
1
K
J
1
1
500
400
300
200
100
LEGEND
100 TOPOGRAPHIC ELEVATION
SUB-BASIN BOUNDARY
DIRECTION OF SURFACE WATER MOVEMENT

Figure 14. Schematic plan view of Plains Regions environment.

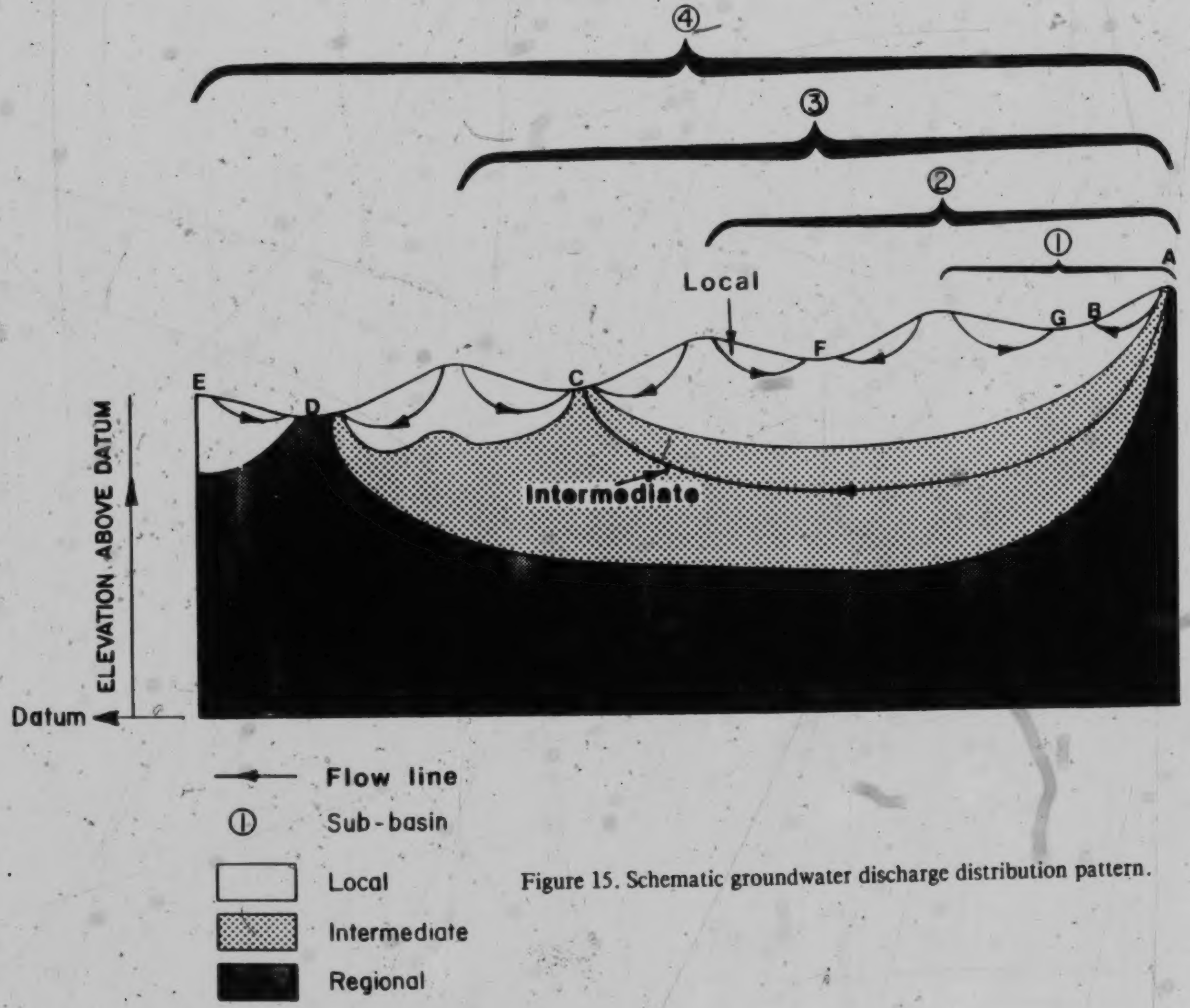

④
③
②
①
A
Local
G
B
F
C
E
D
Intermediate
ELEVATION ABOVE DATUM
Datum
Flow line
① Sub-basin
Local
Intermediate
Regional

Figure 15. Schematic groundwater discharge distribution pattern.

lost partly through a local system at B, partly through an intermediate system at C, and partly through intermediate and regional systems at D. This situation results in less than the amounts of water recharged into each of sub-basins 1, 2 and 3 being discharged within that sub-basin. At the size of basin 4, however, all the water recharged into the system is discharged from it completely. Examination of Figure 15 also indicates that increasingly more water per unit area should be discharged with increasing area of sub-basin up to basin 4 and thereafter, where a relatively constant value could be expected. Two points should be noted: 1. groundwater is lost from sub-basins at high elevation to those at low elevation; 2. the loss of water results in a systematic distribution of regional groundwater discharge that is marked by a baseflow regularity.

If this proposition of systematic groundwater discharge in the Plains Regions is to be usefully employed in regional groundwater development and management, its occurrence must be substantiated and built into a form that is easily applicable to groundwater studies. Furthermore, deviations from the simple picture of the proposed groundwater discharge distribution could be expected to result from changes in hydrogeologic environmental factors other than those considered in the proposition. The remaining sections of this chapter are devoted to the description of the Plains Regions environments in terms of these factors. Following that, in the next two chapters, a detailed theoretical analysis of regional groundwater flow is done, together with a consideration of the most important environmental factors for the purpose of regional groundwater development and management.

## 2.2 Theoretical aspects of characterization

### 2.2.1 The concepts of hydrogeologic environment and groundwater regime

Groundwater studies in general, and the present study in particular, are executed in order to describe, explain, and predict groundwater conditions and related problems. The achievement of these objectives requires that certain properties of groundwater be quantified, for which a knowledge of several components of the groundwater system is necessary. Although some of the components were employed in groundwater studies several decades ago, it was after the development of the solution to the boundary value problem in regional cross section by Tóth (1962) that others were understood in detail.

Tóth (1970) developed a theoretical framework which is capable of identifying, defining and quantifying the significant properties of groundwater, and establishing functional relations between the properties and their controlling components. The framework consists of two sets of physical and chemical parameters; the groundwater regime, and the hydrogeologic environment. The regime parameters, and the elements of the hydrogeologic environment are related by exact or conceptual mathematical equations, hence a knowledge of the environment is used in the calculation of any or all aspects of the groundwater regime (Tóth, 1970). What now follows, is a summary of the concepts of the groundwater regime and the hydrogeologic environment according to Tóth (1970). Only the aspects relevant to this study are discussed, but the interested reader is referred to the paper by Tóth (1970) for a detailed description of these concepts.

#### 2.2.1.1 The groundwater regime

The groundwater regime is the aspect of a given geographic region pertaining to the groundwater and related phenomena. In the present study, a description of the regime requires that the following natural parameters be known quantitatively: 1. amount of water stored in the rocks; 2. the spatial distribution of water movement; 3. the rate of volume discharge or the velocity of flow; and 4. the changes of the above parameters with respect to time.

The amount of water stored in a saturated rock is equal to the pore volume of the rock mass. It is therefore, a function of the geology of the rock.

Groundwater moves along flow paths which are organised in space. The spatial distribution of these flow paths may be termed as a flow pattern, whose knowledge implies that the direction and intensity of groundwater flow at any point in the flow region is known. A groundwater flow pattern is controlled by the configuration of the water table and by the permeability distribution of the rocks; therefore, it is a function of the topography, climate, and geology of the setting. Using numerical simulation methods, a flow pattern of groundwater may be obtained quantitatively by determining the fluid-potential distribution of a region, by applying any of equations (6) through (10) and the proper boundary conditions. The geometry of the flow region and the distribution of rock permeability must be known for this purpose.

Once the fluid-potential distribution of a region is known, the volume discharge or velocity of flow can be evaluated by applying (1). A knowledge of the topographic, climatologic, and geologic conditions is, therefore, necessary for the determination of the volume discharge.

The fourth regime parameter of importance in this study is the system of changes with respect to time of the three regime parameters discussed above. This parameter indicates the type and intensity of the time dependent changes of the regime properties at any point within the flow region, thus, characterizing the variability of the groundwater condition. Since this parameter is dependent on the other regime parameters, it must also be controlled by the topography, climate, and geology of the region.

#### 2.2.1.2 The hydrogeologic environment

The preceding paragraphs identify three main components of the groundwater system: topography, climate, and geology as the major controls on the groundwater regime parameters. A combination of the three components is referred to as the hydrogeologic environment.

The distribution pattern and velocity of groundwater flow is influenced by the relief of the water table in the region. The influence is so well documented (Tóth, 1963; Freeze and Witherspoon, 1967; Powers et al., 1967), that in areas where the water table resembles the surface topography, flow patterns may be calculated by substituting the shape of the land surface for the water table. In a basin with evenly sloping flanks, a unit pattern of regional groundwater flow (Tóth, 1984), is developed in which relative to the water table, groundwater flow is descending, lateral and ascending in the recharge, midline and discharge areas, respectively. The intensity of movement decreases with increasing depth and away from the area of midline; and near the lower corners of the flow field, movement is close to stagnant. Changes in topographic relief result in the modification of flow pattern geometry and flow intensity, primarily producing flow systems of different orders (Tóth, 1963). Flow systems of each order also have the three regimes of recharge, transfer and discharge. The geometry and intensity of the various flow systems is also controlled by other basin parameters, such as, the local and regional slopes,and the width and depth of the basin.

The geology of a region is related to the groundwater regime through the distribution of the porosity and hydraulic conductivity of the rocks. The amount of water present in saturated rocks is directly proportional to the porosity of the rocks. Spatial variations in the hydraulic conductivity of the rock framework, caused by the occurrence of strata, lenses and faults of different permeability, modify the flow pattern and intensity of groundwater flow in the region (Figure 16). The modifying effects of strata and lenses may be great enough as to reverse the recharge-discharge regimes of the flow pattern that would exist otherwise. Climatological factors, such as precipitation, air temperature, evapotranspiration, wind speed, and solar radiation also control some parameters of the groundwater regime. The balance between atmospheric recharge and discharge as controlled by the factors above, determines the configuration of the water table which, in turn, generates the groundwater flow pattern. Whether a region is humid or arid affects the flow rates, and to some extent, the types of flow systems that will exist.

In summary, it should be possible to describe and analyze the groundwater conditions in the Plains Regions of Alberta, from a knowledge of the generating physical environment. The environment has three components, namely: topography, geology, and climate. Each of these components consists of environmental parameters that can be measured, such as actual topographic relief, regional slope, width and depth of region, rock porosity and permeability, precipitation, air temperature, evapotranspiration, wind speed, and solar radiation. Some of these required data on the environmental parameters are available in the form of topographic, geologic, and climatologic maps and reports of various studies in the Plains Regions. It is envisaged that these data can be organised and utilized in the rational categorization of commonly occurring hydrogeologic environments in the Plains Regions. Furthermore, the cause-and-effect relationship which exists between the environment and the groundwater regime is genetic, quantitative, unique, and can be expressed by mathematical equations which are amenable to solution given the boundary conditions. The environmental parameters are described in the next section.

## 2.3 Characteristic environmental parameters of the Plains Regions

It is now understood that a knowledge of the hydrogeologic environmental parameters is the necessary and sufficient condition for evaluating the groundwater regime of a region. For the present purposes, only those parameters which are believed by the author to be most important, will be considered. They include those associated with the region's configuration, viz: topographic relief, regional slope, and depth and width of the region; those associted with

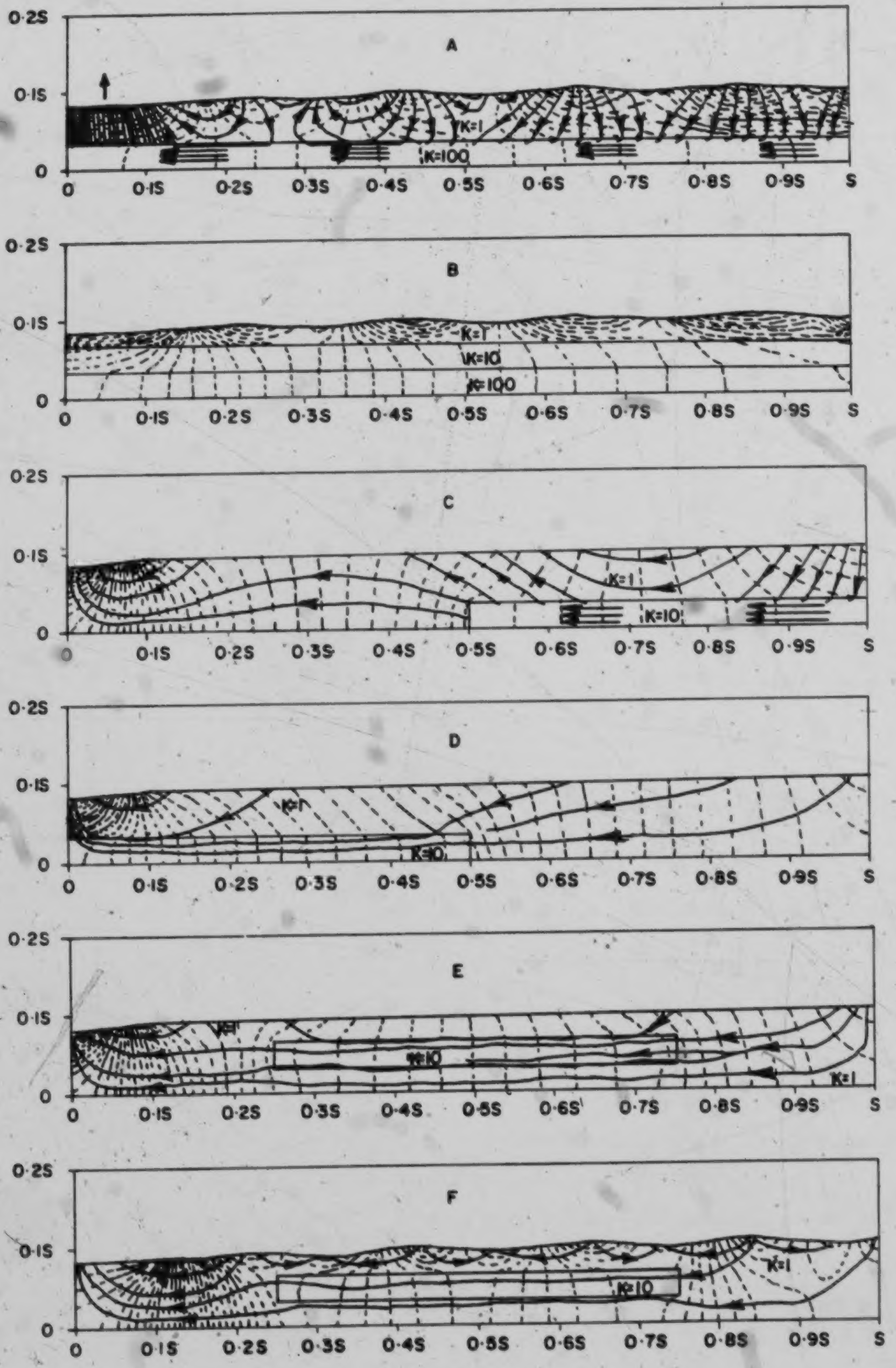

A
K=1
K=100
B
K=1
K=10
K=100
C
K=1
K=10
D
K=1
K=10
E
K=10
K=1
F
K=1
K=10
0.2S
0.1S
0
0.1S
0.2S
0.3S
0.4S
0.5S
0.6S
0.7S
0.8S
0.9S
S

Figure 16. Calculated effects of stratification and lenses on flow distibution (after Freeze and Witherspoon, 1967).

the region's geology, namely: porosity and hydraulic conductivity distribution in the system; and those associated with climate, such as precipitation.

In the Plains Regions of Alberta, the required data on the environmental parameters are present in thousands of maps and reports. The difficulty in using the data, resides in their synthesis and organization in a manner that is easily applicable to groundwater studies in such an extensive area. To this end, the results of the previous baseflow analysis were employed in the selection of the appropriate data. Recalling that a "type environment" in this region has an area of about 6 000 km$^2$ or greater, which implies a width of about 75 km or greater, data from studies of this scale were assumed to be appropriate. Data from more detailed smaller scale studies were used to supplement deductions from the large scale studies where necessary. The Hydrogeological Reconnaissance Map Series (Tóth, 1977) were of invaluable assistance in the description of the Plains Regions environment. Among other aspects, the maps show areally extensive natural hydrogeologic conditions, such as: topographic relief, precipitation, stratigraphy of rocks, lithology of principal aquifers, and hydrostratigraphy. On this basis, the maps have been utilized as basic environmental characterization tools.

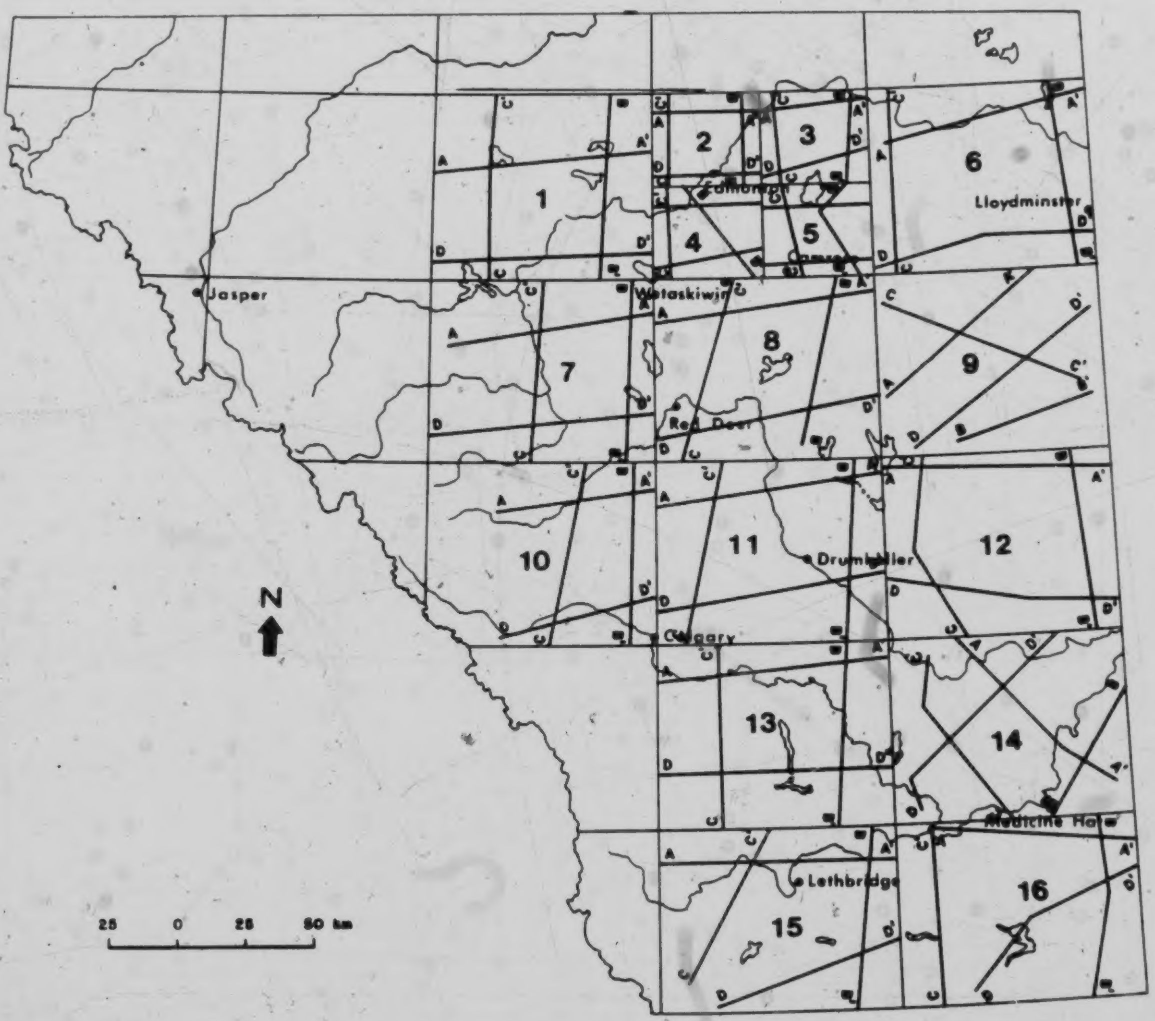
### 2.3.1 "Type environment" of region

A groundwater basin is assumed to represent the type environment in this study, because: 1. regional groundwater development and management in the Plains Regions is done with the basin as a unit; 2. the basin is a convenient unit, and the one that has been utilized by most of the previous studies; 3. the hydrological charcterization of the environment in this study has been carried out with basins as the units of study.

### 2.3.2 Configuration of the basin

#### 2.3.2.1 Depth and width of basin

For practical purposes, a typical basin in the Plains Regions is considered to have a width of at least 75 km, as indicated by the hydrological characterization. In order to determine the contents of every 75 km stretch in the Plains Regions, several geologic cross sections of the area of study, and whose traces are indicated in Figure 17, are presented in Appendix 1. The cross sections are those of previous regional groundwater studies in the area, as summarized in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11
12
13
14
15
16
Edmonton
Camrose
Lloydminster
Jasper
Wetaskiwin
Red Deer
Drumheller
Calgary
Medicine Hat
Lethbridge
N
25 0 25 50 km

Figure 17. Traces of geologic cross sections.

the Hydrogeological Reconnaissance Maps. The smallest depth of regional groundwater study in the area is about 137 m (Appendix 1, plate 4a), whereas, the thickness of sediments is known to exceed 1 830 m in places. For convenience, the depth to width ratio rather than depth or width alone will be analyzed. The depths of 137 m and 1 830 m relative to the width of 75 km, result in depth to width ratios of 0.002 and 0.025, respectively. In the past, basins of smaller widths than 75 km have also been used as study units, and their effects will be evaluated by considering some higher depth to width ratios than estimated here.

#### 2.3.2.2 Regional slope

In the measurement of regional slope, topographic maps having a scale of 1:500 000 and contour interval of 30 m were used. Values of regional slope were measured over distances of 75 km, and then averaged over every one degree of latitude and longitude (Figure 18). The estimated slopes range between 0.0013 and 0.03.

#### 2.3.2.3 Topographic relief

The cross sections in Appendix 1 show topographic profiles, in the regional sense, that can be expected to cover the possible range of profiles in the region of study: In general, elevations vary from less than 520 m above sea level in the valley of the North Saskatchewan River (Appendix 1, plate 6b) to over 1 370 m in the Cypress Hills (Appendix 1, plate 16b). For the present purposes, the two most salient and important characteristics of the topography are: 1. the presence of smooth, featureless land surfaces which rise gradually from a central depression (Appendix 1, plates 2b, 2d, 4a, 4b and 8b), and 2. the complication of these smooth land surfaces by almost regularly occurring alternations of topographic peaks and troughs (Appendix 1, plates 1d, 7a, 7b, 8c, 9a, 9d, 11a, 13a and 14b). Other land surfaces are intermediate between these two end members. For practical purposes, and for simplicity, the first and second end members of the land surface configuration can be repersented by linear and sinusoidal surfaces (Tóth, 1963), respectively, to yield reasonably representative results.

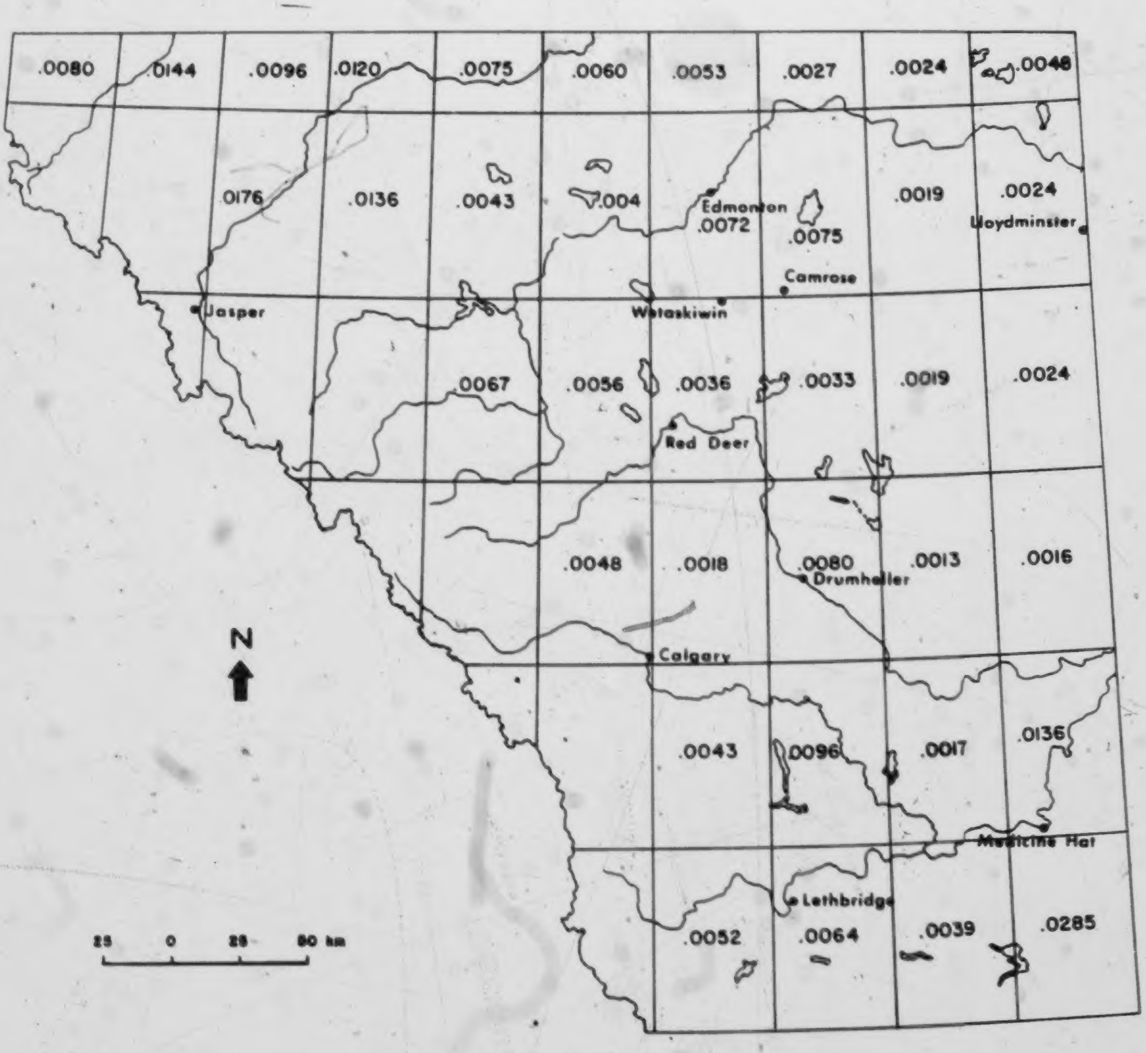

.0080
.0144
.0096
.0120
.0075
.0060
.0053
.0027
.0024
.0048
.0176
.0136
.0043
.004
Edmonton
.0072
.0075
.0019
.0024
Lloydminster
Camrose
Jasper
Wetaskiwin
.0067
.0056
.0036
.0033
.0019
.0024
Red Deer
.0048
.0018
.0080
Drumheller
.0013
.0016
N
Calgary
.0043
.0096
.0017
.0136
Medicine Hat
Lethbridge
25
0
25
50 km
.0052
.0064
.0039
.0285

Figure 18. Regional slopes in Plains Regions of Alberta.

### 2.3.3 Geology of the basin

The Plains Regions of Alberta are underlain by more than 1 830 m of indurated sediments of great diversity which resulted from numerous cycles of marine and fresh-water sedimentation. Four major cycles of sedimentation, corresponding to the Paleozoic (Cambrian), Paleozoic (Devonian-Mississippian), Mesozoic (Cretaceous), and Cenozoic (Quaternary) are recognized (Webb, 1954; Alberta Society of Petroleum Geologist, 1966). The four cycles, underlain unconformably by the crystalline Precambrian basement rocks, are separated by major unconformities representing hiatuses during which subaerial erosion was active. In addition to the major unconformities, minor unconformities, resulting from short time intervals of erosion within each major cycle of sediments, have produced features of hydrologic significance.

In this analysis, because most groundwater studies have been restricted to shallow depths of the sediments, only the stratigraphic section of the Upper Cretaceous and younger sediments is presented in Figure 19. A lithologic evaluation of these sediments usually reveals a sequence of alternating sandstone- and shale-dominated formations.

Most of the previous groundwater studies in the region of interest have considered the sandstone- and shale-dominated geologic units as permeable and impermeable rock units, respectively. Hundreds of hydraulic conductivity and porosity values for each unit are reported in literature, and have been summarized as in Tables 3 and 4 (Tóth, personal communication).

In recent years, however, as a result of increasing development of various quantitative methods for analyzing and interpreting the nature of aquifers and confining beds, it has become evident that the assumption of uniform permeability for all geologic units, based solely on lithologic and paleontologic criteria could lead to error in predicting their true hydrologic behaviour. Other parameters, not commonly utilized, but necessary in the critical definition of rock units in terms of their geohydrologic characteristics, are listed by Maxey (1964). These parameters include, the nature of movement, storage, and release of water in and through various media; the detailed complexities of interstratification, facies, sorting and packing of grains; distribution and degree of interconnection of joints and other fractures; and the degree and depth of weathering and the development of solution channels and caves, both near the present land surface and at levels of former weathering and erosion that are now buried. In this regard, the modeling of the Plains Regions environment on the basis of geologic units is

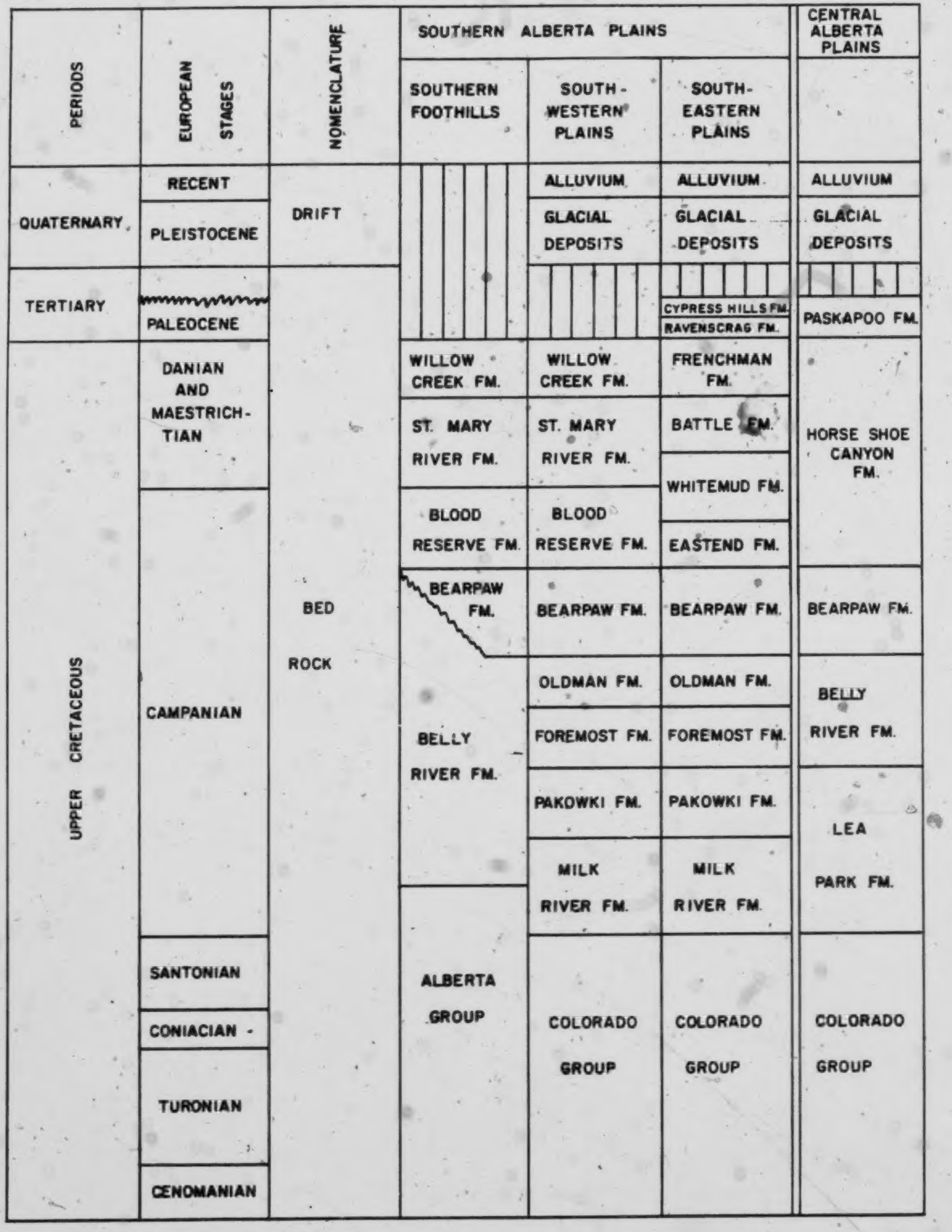

PERIODS
EUROPEAN STAGES
NOMENCLATURE
SOUTHERN ALBERTA PLAINS
CENTRAL ALBERTA PLAINS
SOUTHERN FOOTHILLS
SOUTH-WESTERN PLAINS
SOUTH-EASTERN PLAINS
QUATERNARY
RECENT
PLEISTOCENE
DRIFT
ALLUVIUM
ALLUVIUM
ALLUVIUM
GLACIAL DEPOSITS
GLACIAL DEPOSITS
GLACIAL DEPOSITS
TERTIARY
PALEOCENE
CYPRESS HILLS FM.
RAVENSCRAG FM.
PASKAPOO FM.
UPPER CRETACEOUS
DANIAN AND MAESTRICH-TIAN
CAMPANIAN
SANTONIAN
CONIACIAN
TURONIAN
CENOMANIAN
BED ROCK
WILLOW CREEK FM.
WILLOW CREEK FM.
FRENCHMAN FM.
ST. MARY RIVER FM.
ST. MARY RIVER FM.
BATTLE FM.
WHITEMUD FM.
HORSE SHOE CANYON FM.
BLOOD RESERVE FM.
BLOOD RESERVE FM.
EASTEND FM.
BEARPAW FM.
BEARPAW FM.
BEARPAW FM.
BEARPAW FM.
OLDMAN FM.
OLDMAN FM.
BELLY RIVER FM.
FOREMOST FM.
FOREMOST FM.
BELLY RIVER FM.
PAKOWKI FM.
PAKOWKI FM.
LEA PARK FM.
MILK RIVER FM.
MILK RIVER FM.
ALBERTA GROUP
COLORADO GROUP
COLORADO GROUP
COLORADO GROUP

Figure 19. Stratigraphic section of Upper Cretaceous and younger sediments.

Table 3. Hydraulic conductivity values of commonly occurring formations in Alberta.

| Formations | (K) Hydraulic conductivity (m/s) | | |
|---|---|---|---|
| | Probable range within formation | Probable range of average | Estimated average |
| Till, Lacustrine deposits, Lea Park, Pakowki, Alberta Gp., Wapiabi, Blackstone, Fernie Gp., Colorado Gp, Joli Fou, Labiche, Clearwater, Smoky Gp., Puskwaskau, Kaskapau, Shaftesbury, Low Spirit River, Loon River. | up to<br>$5.6 \times 10^{-9}$ | up to<br>$5.6 \times 10^{-10}$ | up to<br>$2.8 \times 10^{-10}$ |
| Dunvegan, Pouce Coupe, Bearpaw, Aeolian deposits, Cardium, Blairmore Gp., Upper Manville, Top Spirit River. | $5.6 \times 10^{-11}$<br>to<br>$5.6 \times 10^{-9}$ | $5.6 \times 10^{-10}$<br>to<br>$5.6 \times 10^{-9}$ | $2.8 \times 10^{-9}$ |
| Edmonton, St. Mary, Wapiti, Belly River, Foremost, Oldman, Bulmonk, Willow Creek, Lower-Upper Manville. | $5.6 \times 10^{-10}$<br>to<br>$5.6 \times 10^{-7}$ | $5.6 \times 10^{-9}$<br>to<br>$5.6 \times 10^{-8}$ | $2.8 \times 10^{-8}$ |
| Milk River, Ribstone Creek, Lower Birch Lake, Bow Island, Vicking, Top Upper Maville, Bad Heart, Bluesby, Nikanamin. | $5.6 \times 10^{-9}$<br>to<br>$5.6 \times 10^{-6}$ | $5.6 \times 10^{-8}$<br>to<br>$5.6 \times 10^{-7}$ | $2.8 \times 10^{-7}$ |
| Paskapoo, Lower Manville, Ellerlie, Deville, Grand Rapids, Bullhead Gp., Codomin, Gelling, Pelican, Top Nikanamin. | $5.6 \times 10^{-8}$<br>to<br>$5.6 \times 10^{-5}$ | $5.6 \times 10^{-7}$<br>to<br>$5.6 \times 10^{-6}$ | $2.8 \times 10^{-6}$ |
| McMurray, Porcupine Hills. | | $5.6 \times 10^{-6}$<br>to<br>$5.6 \times 10^{-5}$ | $2.8 \times 10^{-5}$ |
| Recent alluvial terraces, sands and gravells of buried valleys, sands and gravells of glacial meltwaters. | $5.6 \times 10^{-6}$<br>to<br>$5.6 \times 10^{-2}$ | $2.8 \times 10^{-5}$<br>to<br>$2.8 \times 10^{-3}$ | $5.6 \times 10^{-4}$ |

Source - Tóth, personal communication

Table 4. Porosity values of some formations in Alberta.

| Formation | Lithology | Porosity | | |
|---|---|---|---|---|
| | | Type | Range % | Average % |
| Wapiti | sandstone + shale | intergranular | 34.2 - 11.2 | 22.0 |
| Milk River | | intergranular | 10.0 - ? | 10.0 |
| Alluvium River Bank | | intergranular | 35.0 - ? | 35.0 |
| Ellerslie | dolomite | vuggy + fracture | 30.0 - 12.5 | 20.0 |
| Nisku | dolomite | vuggy + fracture | 10.9 - 3.0 | 5.3 |
| Leduc | dolomite | vuggy + fracture | 10.8 - 3.0 | 7.1 |
| Cardium | sandstone | | 18.3 - 7.8 | 12.7 |
| Sunburst | sandstone | | 23.5 - ? | 23.5 |
| Wainwright | sandstone | intergranular | 31.0 - 25.0 | 26.7 |
| Cessford | | | 27.1 - 23.0 | 24.5 |
| Vicking | | vuggy | 23.0 - 7.9 | 15.4 |
| Fish Sc. | sandstone | | 23.0 - ? | 23.0 |
| Wabamun | | | 3.5 - 3.3 | 3.4 |
| Colony | | | 7.2 - ? | 7.2 |
| Swan Hills | | | 9.0 - 6.6 | 8.6 |
| Conrad | sandstone | | 17.4 - ? | 17.4 |
| Rundle | | | 10.0 - 7.0 | 8.5 |
| Banff | | | 11.0 - ? | 11.0 |
| Glauc. | sandstone | | 17.0 - 14.9 | 15.9 |
| Gilwood | sandstone | | 14.3 - ? | 14.3 |
| Elkton | | | 13.8 - 5.5 | 9.9 |
| Pekisko | | | 13.3 - ? | 13.3 |
| Ostracod | | | 20.8 - ? | 20.8 |
| Granite wash | | | 13.0 - 11.5 | 12.2 |
| Belly River | | | 19.0 - 15.0 | 17.3 |
| Arcs | | | 15.0 - 8.0 | 11.5 |
| Moulton | sandstone | | 17.5 - ? | 17.5 |
| Cooking Lake | | | 5.0 - ? | 5.0 |
| Todd-gray | | | 15.0 - ? | 15.0 |
| Cutbank | sandstone | | 21.0 - 17.3 | 19.2 |
| Turner River | | | 12.8 - 9.8 | 11.3 |

Source - Tóth, personal communication.

particularly questionable, due to the conditions of deposition of the units. It is presumed that a description based on hydrogeologic units will better represent the environment as required for the present analysis.

### 2.3.3.1 Hydrogeologic units

A hydrogeologic unit is a body of rock that can be distinguished as either water bearing or water retarding relative to adjacent rocks. The sequence of sedimentation in the western Canada sedimentary basin, of which the Plains Regions is part seems to have taken one of two directions, resulting in distinguishable hydrogeologic units. Prior to the Cenozoic era, the first direction, marked by the encroachment of the sea on land, resulted in the accumulation of regionally extensive marine shales mixed with local sandstones of such sub-environments as deltas. Marine regressions often followed the transgressions, with the deposition of continental sandstones intercalated by shales and siltstones of lower energy sub-environments. During the Quaternary, a series of glacial advances and retreats paralleled the pre-Cenozoic marine transgressions and regressions, respectively (Stalker and Harrison, 1977; Rutter, 1980). Whereas the glacial advances were marked by till deposits of low permeability, outwash sands and gravels of meltwater channels, and loesses registered interglacial retreats. A further complication in the relationship of the high- to low-permeability hydrogeologic units is caused by erosion during major and minor hiatuses. Subaerial erosion produced increased permeabilities and porosities within and between geologic units relative to the surrounding strata.

The resultant effect of the sequence of events is the existence of hydrogeologic units of high permeability (aquifers) or low permeability (aquitards), which are independent of the primarily deposited geologic units (Figure 20).

In order to establish the nature of occurrence of the aquifers and aquitards in the present study, the hydrostratigraphic cross sections of the Hydrogeological Reconnaissance Maps (Appendix 1) were used as the basic tools. Originally, the hydrostratigraphy is described in terms of twenty-year safe yield ($Q_{20}$), the constant rate at which a well can produce continuously so that at the end of twenty years, the water level will be drawn down to the top of the producing layer. Safe yields ranging from less than $7.6 \times 10^{-5}$ to over $3.8 \times 10^{-3}$ m$^3$/s have been estimated for various hydrogeologic units in the region of study, through the equation

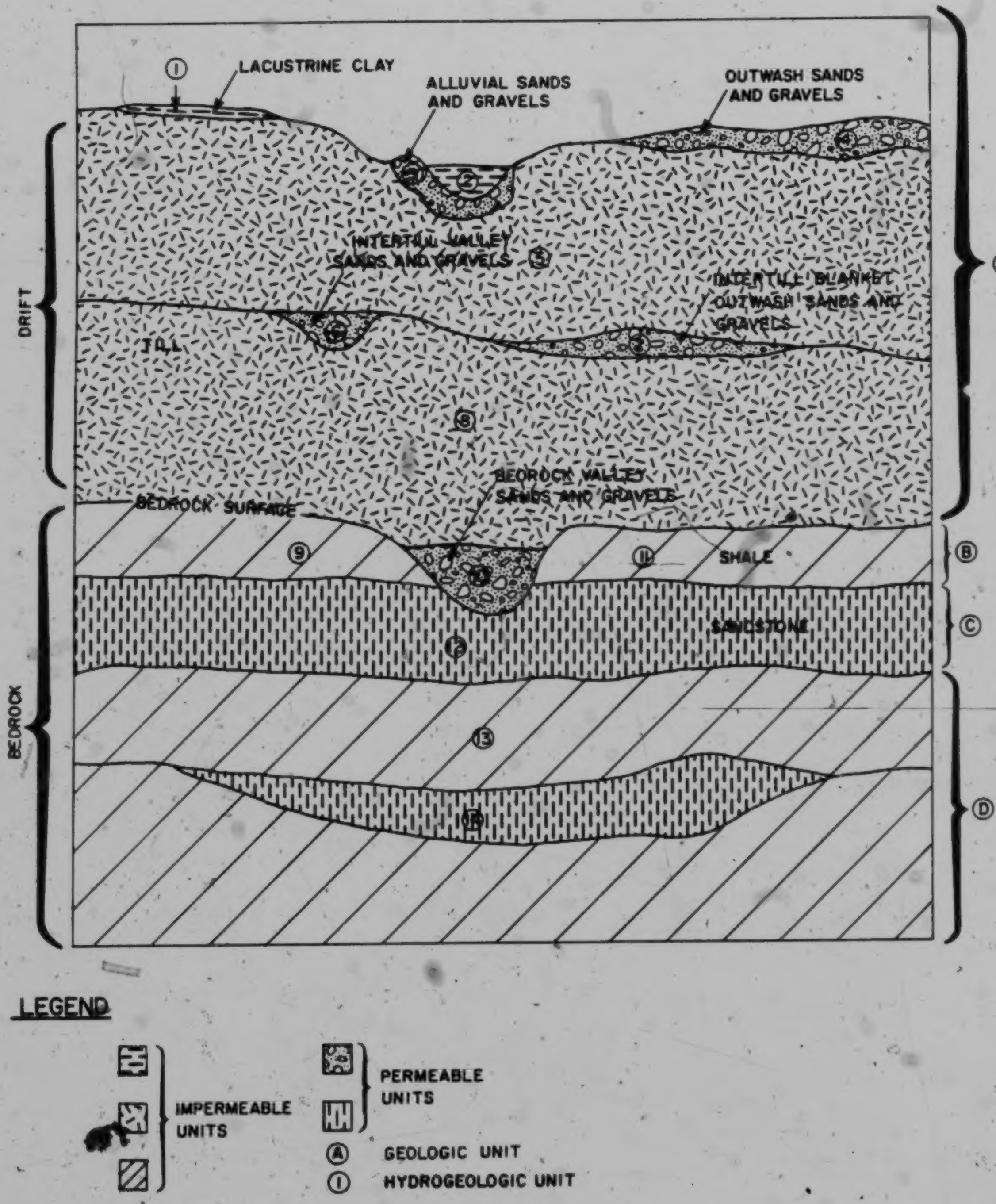

①
LACUSTRINE CLAY
ALLUVIAL SANDS AND GRAVELS
OUTWASH SANDS AND GRAVELS
②
③
④
INTERTILL VALLEY SANDS AND GRAVELS
⑤
INTERTILL BLANKET OUTWASH SANDS AND GRAVELS
⑥
⑦
TILL
⑧
DRIFT
BEDROCK VALLEY SANDS AND GRAVELS
BEDROCK SURFACE
⑨
⑩
⑪
SHALE
SANDSTONE
⑫
⑬
⑭
BEDROCK
Ⓐ
Ⓑ
Ⓒ
Ⓓ
LEGEND
IMPERMEABLE UNITS
PERMEABLE UNITS
Ⓐ GEOLOGIC UNIT
① HYDROGEOLOGIC UNIT

Figure 20. Schematic illustration of geologic and hydrogeologic units.

(Tóth, 1966b):

$$Q_{20} = \frac{TH \times 0.7}{2110} = \frac{KbH \times 0.7}{2110} \qquad (16)$$

where: $Q_{20}$ = twenty-year safe yield (igpm);
$T$ = transmissivity of the unit (igpd/ft);
$K$ = hydraulic conductivity of the unit (igpd/ft$^2$);
$b$ = thickness of the unit (ft);
$H$ = available drawdown in the unit (ft).

A careful examination of all the reports reveals that hydraulic conductivity is the major controlling factor on the values of safe yield estimated for the units through (16).

The separation of the aquifers from the aquitards required the establishment of a standard. Thus, an aquifer was defined further as a rock unit which is permeable enough to yield economic quantities of water to wells (Freeze and Cherry, 1979), and an arbitrary yield of $2.0\mathrm{x}10^{-3}$ m$^3$/s was selected to mark the boundary between economic and non-economic rates. This rate would contribute significantly to water needs in a little settlement in the Plains Regions. An aquifer, therefore, will yield $2.0\mathrm{x}10^{-3}$ m$^3$/s or more, whereas an aquitard yields less than this value.

Using these criteria, the mode of occurrence of aquifers and aquitards in relation to geologic units in the Plains Regions is displayed in the cross sections of Appendix 1. Recognizing that the type environment has a width of 75 km, the cross sections may be used in summarizing the mode of occurrence of aquifers and aquitards in a general way, as follows: 1. aquifers and aquitards are independent of the geologic units; 2. no aquifer exists in some regions (Plates 2d, 4c, 14c and 16b); 3. where they occur, aquifers vary in: (a) sizes from areally extensive layers (Plates 1d, 3a, 6a, 7a, 7c, 8c, 10a and 15b) to restricted lenses of various forms (Plates 1a, 1b, 2b, 2c, 3b, 3c, 3d, 4a, 4b, 4d, 5a, 5c, 5d, 6b, 6c, 6d, 7b, 8a, 8b, 8d, 9a, 9b, 9c, 9d, 10b, 10c, 11a, 11c, 11d, 13b, 13c, 13d, 14a, 14b, 14d, 15a, 15b, 15c, 15d, 16a, 16c and 16d); (b) position from deep (Plates 2a, 2b, 3a, 3c, 3d, 5a, 5b, 5d, 15a and 15b) to shallow (Plates 1a, 1b, 1c, 2b, 2c,3a, 3b, 3c, 3d, 4a, 4b, 4d, 5a, 5c, 5d, 6a, 6b, 6c, 6d, 7a, 7b, 7c, 7d, 8a, 8b, 8c, 8d, 9a, 9b, 9c, 9d, 10a, 10b, 10c, 11a, 11b, 11c,11d, 13a, 13b, 13c, 13d, 14a, 14b, 14d, 15a, 15b, 15c, 15d, 16a, 16c, and 16d), and from high surface topographic elevations (recharge areas, Plates 1b, 1d, 3c, 3d, 4b, 5c, 6a, 6c, 6d, 7a, 7b, 7c, 7d, 8c, 8d, 13a, 13b, 13c and 15d) to low surface topographic elevations (discharge areas, Plates 1a, 1b, 1c, 2b, 2c, 3a, 3c, 5a, 5c, 6c, 7b, 7c, 7d, 8b, 8c, 9a, 9c, 10b, 10c, 13a, 13c, 14a, 14b, 15b, 15c, 15d and 16a).

Although the discrimination of aquifers from aquitards was based on yield, the yield is controlled mainly by the hydraulic conductivities of the units, therefore, high hydraulic conductivities (Table 3) can be used to represent aquifers, while low hydraulic conductivities are assigned to aquitards. It is noteworthy also, that only an arbitrary and convenient yield has been employed in the separation of aquifers and aquitards. In practice however, groundwater yield has economic value that is proportionate to the use for which it is needed. Thus, some units which are termed as aquitards herein, might serve as aquifers under certain circumstances. Hydraulic conductivities range from $5.6 \times 10^{-4}$ to $2.8 \times 10^{-10}$ m/s (Table 3), for hydrogeologic units in the Plains Regions environments, and will be considered sytematically in the present study. Similarly, representative values of porosity (Table 4) will be employed.

### 2.3.4 Climate of the basin

Climatic elements, such as precipitation, temperature, wind speed, solar radiation, actual evapotranspiration and so on, all control the relief of the water table, the generator of the groundwater flow pattern. Precipitation appears to be the key climatic factor because, it is the main source of input of water into the groundwater system. Other factors act to modify the input from precipitation, both temporally and spatially. For instance, temperature, wind speed, and solar radiation regulate the actual evapotranspiration, which is a major determinant of actual infiltration from precipitation into the subsurface. Also, temperature determines the length of, and period of frost free days, during which groundwater recharge is possible. Since it is impracticable to evaluate the effects of every climatic factor in necessary detail in any one study, only precipitation will be considered in the present study.

#### 2.3.4.1 Precipitation

Over one hundred precipitation stations exist in the Plains Regions for which daily, monthly and annual maximum, minimum and mean precipitation values of a sufficiently long period of time are recorded. An annual precipitation map of the area of study (Figure 21) was

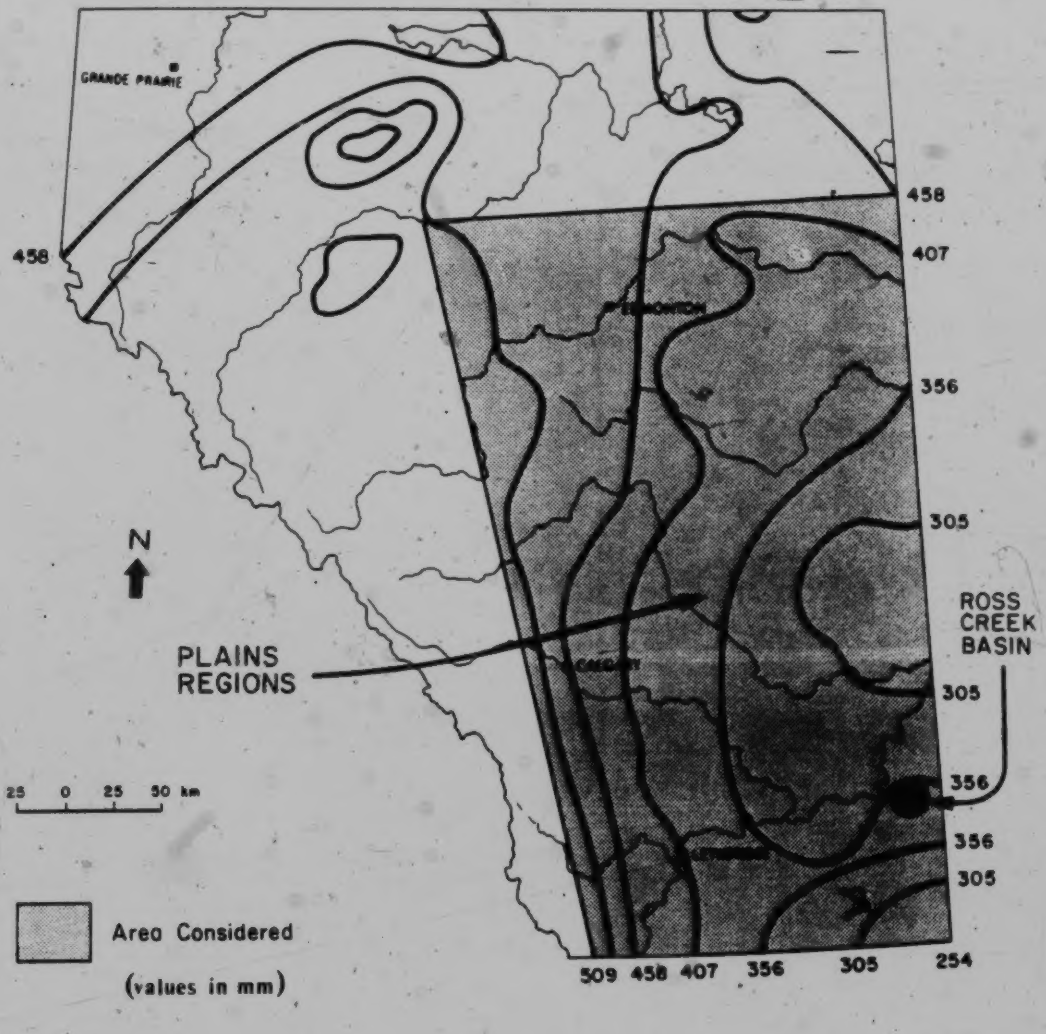

GRANDE PRAIRIE
EDMONTON
CALGARY
LETHBRIDGE
458
458
407
356
305
305
356
356
305
509
458
407
356
305
254
ROSS CREEK BASIN
PLAINS REGIONS
N
25 0 25 50 km
Area Considered
(values in mm)

Figure 21. Average annual precipitation map (after Longley, 1968)

used in estimating possible precipitation rates. Although this map was constructed from precipitation normals of 1931 to 1960 originally, it was updated with data from the Atlas of Alberta (Atlas of Alberta, 1968) and precipitation normals of 1951 to 1980 (Environment Canada, 1985). Precipitation is lowest in the southeastern corner and highest in the western margins of the Plains, corresponding to average rates of $7.0 \times 10^{-4}$ and $1.4 \times 10^{-3}$ m/day.

These estimates of precipitation rates are based on the assumption that annual precipitation is uniformly distributed. However, annual precipitation may be concentrated in six, four, three, or even one month. Assuming a special case in which all the 509 mm of precipitation (Figure 21) occurs in only one month of the year along the western margins of the Plains, a maximum precipitation rate of $1.7 \times 10^{-2}$ m/day should be recorded. On the other hand, precipitation may be negligible in parts of the Plains in some years, resulting in rates of near zero. In the transient analysis of the groundwater system in chapter 4, precipitation rates ranging from 0.0 to $1.7 \times 10^{-2}$ m/day will be considered.

### 2.3.5 Water use

Water has value by virtue of its use. Therefore, an evaluation of regional groundwater flow for basin management will be incomplete without a consideration of the use for which the water is needed.

Water in the Plains Regions is used for the following major purposes (Water Resources Act, 1970): agriculture, diversion, domestic, industrial, irrigation, municipal, power, storage, residential, and other purposes. With minor exceptions, the Water Resources Act requires that all water taken from any part of the Plains be licensed by Alberta Environment. Each license issued specifies the amount of water use permitted, the location of the withdrawal, and how the water will be used. For a number of reasons, the license file does not indicate the amount of water being withdrawn, accurately. First, a client may obtain a license for more water than actually is needed. Second, some water users fail to obtain the required licenses. Third, water license information is inaccurate in a few instances. Fourth, water licenses indicate permitted withdrawals but give no evidence on how much water is returned to the system. Despite these weaknesses, the active water license file is sufficiently reliable to permit generalizations about the nature of water use as required in the present analysis.

Records from the active water license file for 1984 have been summarized (Table 5). In order to obtain the unit rate of water use, it was necessary to construct a monthly water demand matrix for the individual types of water use (Table 6). The number of days during which water is demanded for any specific purpose in the year is determined from Table 6. This value, together with the number of unit applicants are used as divisors to the total annual volume for the calculation of the rates on Table 5.

For the present purposes, a relative classification of all the water uses in terms of volume and rate was necessary. Consequently, an arbitrary scale for both volume and rate was set as follows: less than 100: Low (L); between 100 and 1 000: moderate (M); and greater than 1 000: high (H). Using this classification scheme, all the water use types in the region of study have been classified as indicated in Table 7. The table reveals that water use types vary from those requiring low volume and low rate (LL) of supply to those requiring high volume and high rate (HH) as exemplified by domestic and irrigation uses, respectively. It is presumed that the development of a Plains Regions basin to satisfy two different uses, such as domestic and irrigation requires different strategies of well location within the basin.

### 2.3.6 Categories of Plains Regions environments

It is now possible to categorize the expected environments in the Plains Regions from our knowledge of the hydrogeologic parameters in the preceding sections. Only by categorizing environments as complex as the Plains Regions so that general relationships of groups of environments can be seen, will it be possible to attack a systematic evaluation of the environments for groundwater management purposes. The purpose of classifying is to simplify the structure and relationship of the environments so that general statements can be made about the classes of the environments (Sokal, 1974).

First, it is necessary to summarize the hydrogeologic parameters in an applicable manner. Whereas some factors are major or primary, others are minor or secondary (Table 8). While the primary factors exert primary controls of the groundwater pattern, the secondary factors shed secondary effects on the primary flow pattern. A hierarchy of two orders of basic or primary environments (basins) was established for the area on the basis of the primary parameters on Table 8.

Table 5. Provincial statistics of water demands.

| Use (statutory) | No of unit applicants | Total annual volume (millions of m³) | Rate per unit (m³/day) |
|---|---|---|---|
| AGR | 666 | 21.9 | 900.8 |
| DIV | 1 351 | 10.8 | 21.9 |
| DOM | 6 224 | 72.5 | 31.9 |
| IND | 1 089 | 2 690.3 | 6 771.1 |
| IRR | 2 558 | 2 835.9 | 6 075.8 |
| MUN | 987 | 705.2 | 1 957.5 |
| POW | 9 | 252.8 | 76 943.2 |
| STR | 633 | 233.0 | 1 008.4 |
| OTH | 182 | 91.5 | 1 376.9 |
| *RES | 2 349 000 | 164.4 | 0.2 |

* estimated from known per capita rate and population.

Table 6. Monthly water demand matrix for individual type of water use.

| | Use | Jan | Feb | Mar | Apr | May | Jun | Jul | Aug | Sep | Oct | Nov | Dec |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | RESIDENTIAL | 7.5 | 7.2 | 7.3 | 7.7 | 8.7 | 9.4 | 9.5 | 9.6 | 9.1 | 8.4 | 8.0 | 7.8 |
| | MUNICIPAL | 7.5 | 7.2 | 7.3 | 7.7 | 8.7 | 9.4 | 9.5 | 9.6 | 9.1 | 8.4 | 8.0 | 7.8 |
| | THERMAL POWER GENERATION | 9.0 | 9.0 | 6.3 | 6.3 | 6.4 | 9.0 | 9.0 | 9.0 | 9.0 | 9.0 | 9.0 | 9.0 |
| | OILFIELD ACTIVITIES | 8.3 | 8.3 | 8.3 | 8.3 | 8.3 | 8.3 | 8.3 | 8.3 | 8.3 | 8.3 | 8.3 | 8.3 |
| IND | Petroleum refining | 7.2 | 7.2 | 8.1 | 7.4 | 8.5 | 9.4 | 9.2 | 9.4 | 9.2 | 8.6 | 7.7 | 7.6 |
| | Petrochemicals | 7.8 | 7.1 | 8.3 | 8.3 | 8.9 | 7.5 | 9.5 | 9.1 | 8.8 | 8.7 | 7.9 | 8.2 |
| | Metal Processing | 8.4 | 7.3 | 9.0 | 8.6 | 9.2 | 8.8 | 9.0 | 9.2 | 7.4 | 7.8 | 7.3 | 7.8 |
| | Food and Beverages | 10.8 | 7.4 | 7.4 | 7.5 | 7.8 | 8.1 | 8.7 | 9.1 | 8.7 | 8.4 | 8.3 | 8.1 |
| | Other | 7.7 | 7.6 | 7.5 | 8.1 | 8.5 | 8.8 | 9.1 | 9.8 | 9.0 | 8.7 | 7.8 | 7.4 |
| AGR | Stock watering | 7.8 | 7.6 | 7.5 | 8.1 | 8.6 | 8.7 | 9.0 | 9.7 | 9.1 | 8.7 | 7.7 | 7.5 |
| | Feedlot | 8.3 | 8.3 | 8.5 | 8.5 | 8.5 | 8.0 | 8.0 | 8.0 | 8.0 | 8.3 | 8.3 | 8.3 |
| | Water fowl propagation | 0.0 | 0.0 | 0.0 | 7.6 | 15.4 | 15.4 | 15.4 | 15.4 | 15.4 | 15.4 | 0.0 | 0.0 |
| | IRRIGATION | 0.0 | 0.0 | 0.0 | 0.0 | 9.0 | 21.0 | 32.0 | 22.0 | 12.0 | 4.0 | 0.0 | 0.0 |
| | FOREST RELATED ACTIVITIES | 8.5 | 8.5 | 7.8 | 7.8 | 8.5 | 8.5 | 8.5 | 8.5 | 8.5 | 8.5 | 8.5 | 8.5 |
| | RECREATION | 7.1 | 7.1 | 7.1 | 0.0 | 7.2 | 14.3 | 14.3 | 14.3 | 14.3 | 7.2 | 0.0 | 7.1 |

(All values in percentage of annual total)
Source - Canadian Resourcecon ltd., (1982); H. Weaver, P. Jackson, W. Wenkel (personal communication).

Table 7. Relative classification of water use.

| Use | | Total annu. vol. | Rate per unit | Class |
|---|---|---|---|---|
| RESIDENTIAL | | M | L | ML |
| MUNICIPAL | | M | H | MH |
| THERMAL POWER GENERATION | | M | H | MH |
| OILFIELD ACTIVITIES | | L | H | LH |
| | Petroleum refining | H | H | HH |
| I | Petrochemicals | H | H | HH |
| N | Food and Beverages | H | H | HH |
| D | Other | H | H | HH |
| A | Stock watering | L | M | LM |
| G | Feedlot | L | M | LM |
| R | Water fowl propagation | L | M | LM |
| IRRIGATION | | H | H | HH |
| FOREST RELATED ACTIVITIES | | L | H | LH |
| RECREATION | | H | L | HL |
| DOMESTIC | | L | L | LL |

L=LOW M=MODERATE H=HIGH

Table 8. Summary of environmental parameters.

| Environmental factor | Status | Parameter : | Variation |
|---|---|---|---|
| Topography | Primary | Topographic relief : | simple (linear) or complex (sinusoidal) |
| | Secondary | Depth/width ratio : | deep or shallow |
| | | Regional slope : | high or low |
| Geology | Primary | Rock framework : | homogeneous or heterogeneous |
| | | (if heterogeneous) : | extensive aquifer or restricted aquifer |
| | Secondary | Position of aquifer : | top or bottom of system; in recharge, midline or dicharge area |
| | | Hydraulic conductivity : | high or low |
| Climate | Primary | Climatic control on water table: | assumed similar to topographic relief in all cases |
| | Secondary | Precipitation rate : | high or low; constant or varying |

Based on topographic relief, the second-order grouping separates all the primary environments into categories B of simple linearly sloping surface and A of complex sinusoidal surface. On the basis of rock framework, each of categories A and B is subdivided into homogeneous (AH and BH), and heterogeneous (AT and BT) subgroups, in the first-order grouping. Furthermore, subgroups AT and BT are separated into extensive (layer) aquifer (ATE) and restricted (lens) aquifer (BTR) systems on the basis of aquifer sizes.

In total, the six basic or primary environments are shown in Figure 22. Together with each of these primary environments, a systematic consideration of each set of the secondary parameters produces what is termed as a "secondary environment" in this study. Figure 23 illustrates the sequence through which a secondary environment may be obtained. Numerical values characterizing the secondary parameters are obtainable from the preceding sections. It is clear from Figure 23 that numerous secondary environments than can be analyzed in any one study due to limitations in money and time, will result from various combinations of the primary environments and the secondary parameters. For this reason, only few secondary environments will be selected to demonstrate the major objectives of this study in the chapters that follow.

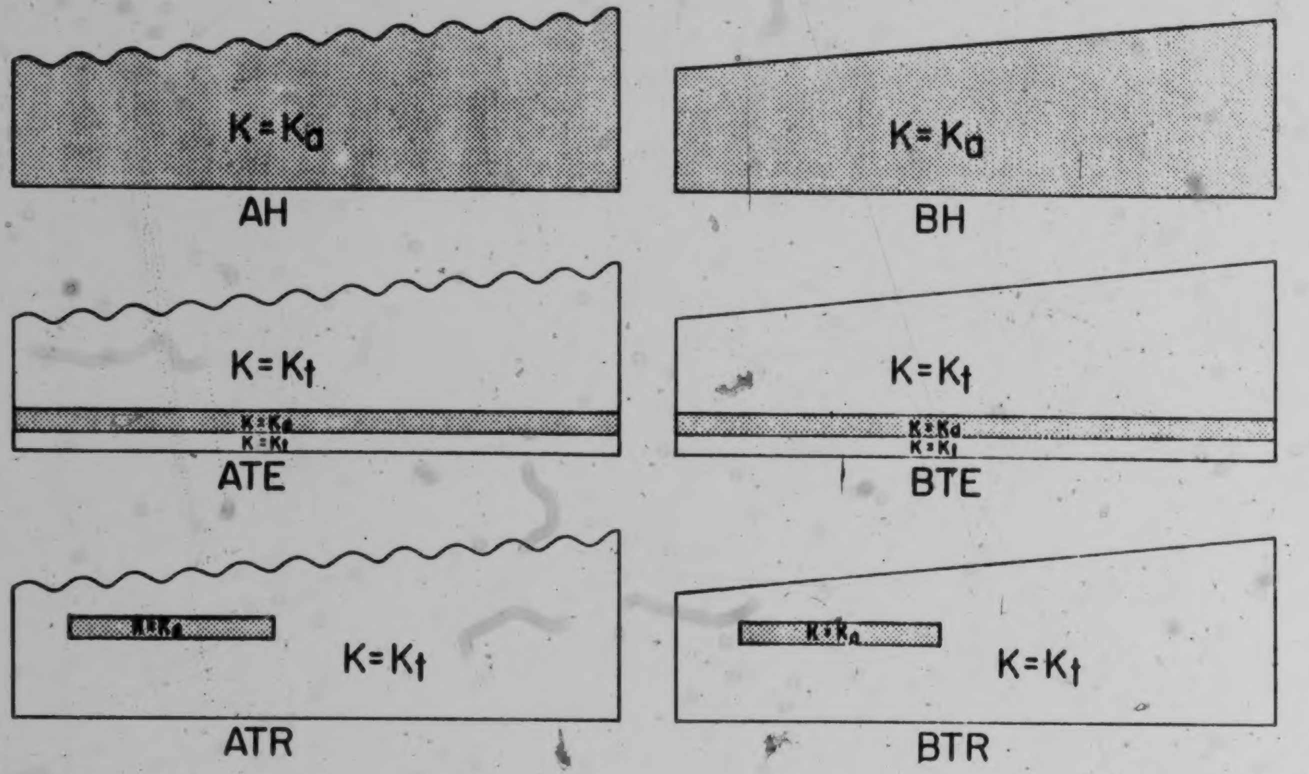

K=Ka
AH
K=Ka
BH
K=Kt
K=Ka
K=Kt
ATE
K=Kt
K=Ka
K=Kt
BTE
K=Ka
K=Kt
ATR
K=Ka
K=Kt
BTR

Figure 22. Primary type environments of the Plains Regions.

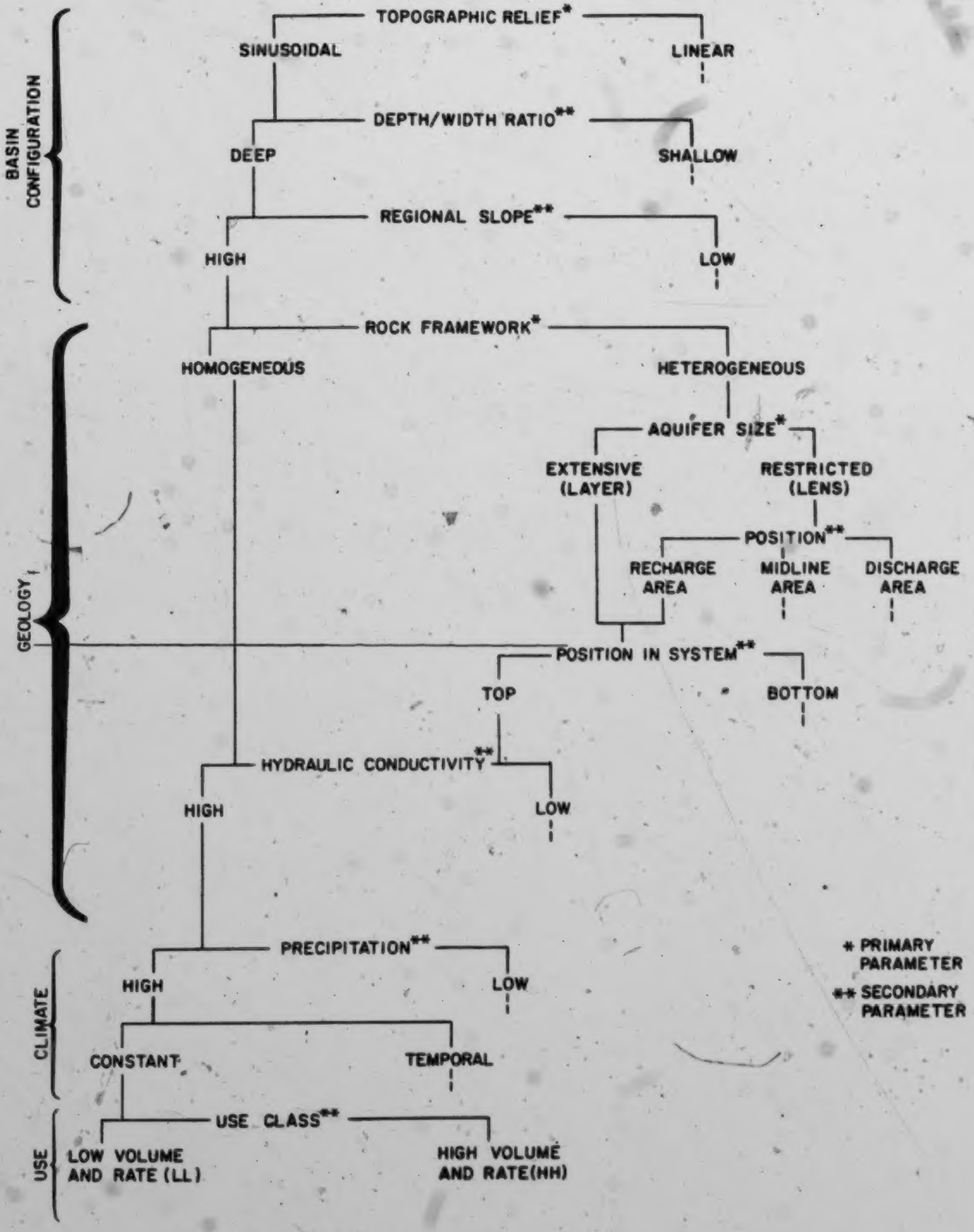

BASIN CONFIGURATION
TOPOGRAPHIC RELIEF*
SINUSOIDAL
LINEAR
DEPTH/WIDTH RATIO**
DEEP
SHALLOW
REGIONAL SLOPE**
HIGH
LOW
GEOLOGY
ROCK FRAMEWORK*
HOMOGENEOUS
HETEROGENEOUS
AQUIFER SIZE*
EXTENSIVE (LAYER)
RESTRICTED (LENS)
POSITION**
RECHARGE AREA
MIDLINE AREA
DISCHARGE AREA
POSITION IN SYSTEM**
TOP
BOTTOM
HYDRAULIC CONDUCTIVITY**
HIGH
LOW
CLIMATE
PRECIPITATION**
HIGH
LOW
CONSTANT
TEMPORAL
USE
USE CLASS**
LOW VOLUME AND RATE (LL)
HIGH VOLUME AND RATE(HH)
* PRIMARY PARAMETER
** SECONDARY PARAMETER

Figure 23. Graphic illustration of parameters of the environment.

# 3. THEORETICAL EVALUATION OF STEADY-STATE REGIONAL GROUNDWATER FLOW IN THE PLAINS REGIONS ENVIRONMENT

## 3.1 General statement

The baseflow regularity discussed in the previous chapter is suggestive of a systematic distribution of regional groundwater discharge in the Plains Regions of Alberta. If one accepts the basic water balance law that groundwater discharge equals groundwater recharge (Meyboom, 1967), and that this "flow through" equals the sustainable basin yield of a virgin basin (Freeze, 1968; Tóth, 1968, 1982), the recognition of a regularity which describes its distribution in a region might be useful in groundwater studies. In section 2.1.2.3, a hypothesis was proposed to explain the possible origin of the baseflow regularity. The verification of such a hypothesis lies in the ability to reproduce the regularity according to the concepts of regional gravity flow of groundwater. Such a duplication may, if successful, in addition to explaining the regularity, contribute to a better understanding of the nature of groundwater movement and its distribution in the Plains Regions.

The equations of groundwater flow and their application in groundwater studies are well documented in literature, and were reviewed in chapter one. The hydrogeologic environmental factors, that is, the generators of the groundwater regime, are known and can be quantified as described in chapter two. The only apparent problem is that of the existence of hundreds of representative environments in the Plains Regions, which has also been simplified by categorizing the environments. The logical next step in defining groundwater movement and its distribution in the Plains is to use numerical modeling to examine detailed patterns of groundwater flow through typical environments.

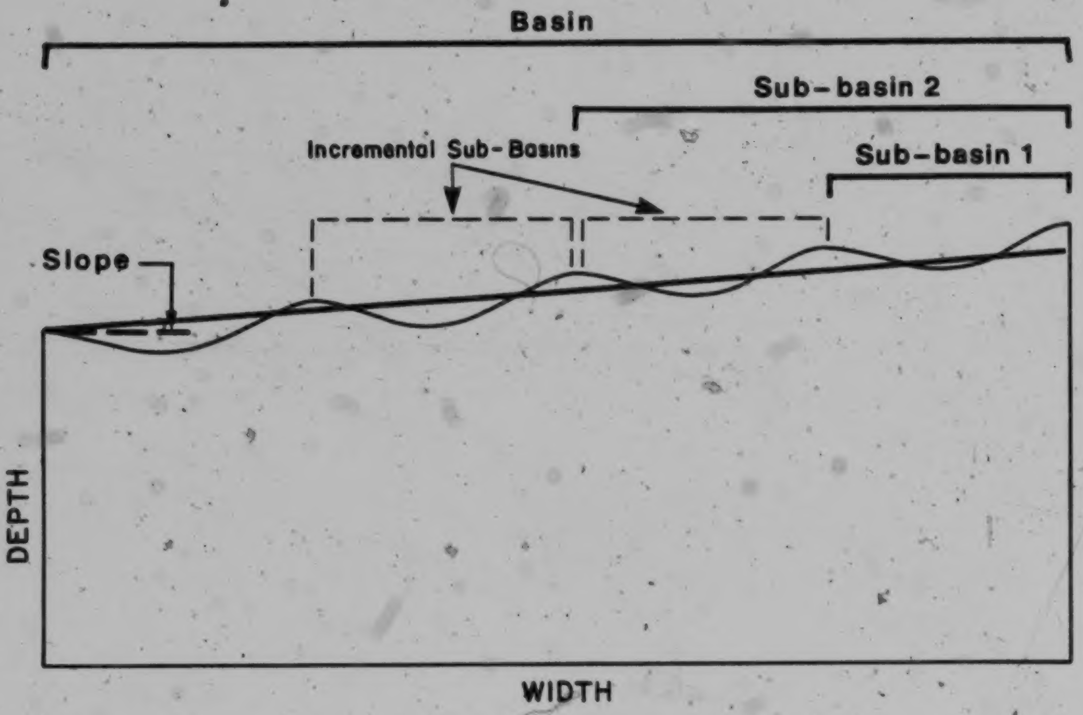
In this chapter, the steady-state flow of groundwater through type Plains Regions environments is analyzed by numerical simulation in order to: 1. reproduce the empirically determined baseflow regularity; 2. determine the controlling functional relations between environmental parameters and the baseflow curve; and 3. examine if recognition of these relations might be usefully employed in the present and future regional groundwater resources development and management studies.

### 3.2 Formulation of problem

The potential field defining the flow system results from the inter-relation of three governing factors: 1. the shape of the region in which the potential field is defined; 2. the existing boundary conditions, and 3. the partial differential equation which describes the flow.

In this study, only two-dimensional, cross-sectional systems are considered, thus only the x- and z-directions are of interest, the y-dimension being infinite. In particular, the AH category (Figure 22) of Plains Regions environments is taken as the basic unit of analysis in this chapter. The area is bounded on the surface by two water divides, one of which occurs at a higher topographic elevation relative to the other. A horizontal impermeable boundary occurs at the base of the domain, whereas, one vertical impermeable boundary extends downward from each surface divide to the base. The justification for the use of such boundaries may be found in the papers by Tóth (1962, 1963), and Freeze and Witherspoon (1968). For reasons discussed previously, a sinusoidal shape was chosen for the surface topography, the highs and lows of which are thought to be representative of the divides and valleys of the sub-basins within the main basin (Figure 24). Basin numbers are also indicated (Figure 25), with basin sizes increasing from high to low topographic elevation, simulating little creeks occurring at high topographic elevations and forming tributaries to streams at intermediate elevations, which then run into rivers at low elevations, as illustrated in Figure 14. In spite of this presumably common situation, it is also possible to find little creeks at low elevations with rivers occurring at relatively higher elevations. A consideration of this case in a later section, together with the former case are, therefore, in agreement with natural laws. Throughout the analysis in this chapter, the water table is at steady state (Freeze, 1966), and it is assumed to coincide with the surface topography (Jones, 1960; Meyboom, 1961b; Farvolden, 1961; Tóth, 1962). The use of the AH environment implies homogeneous conditions. However, the effects of layering and lensing (aquifers) will be evaluated through the introduction of the ATE and ATR environments, respectively.

For a solution to the problem, the boundary conditions above must be defined mathematically (Figure 25). This figure represents an x-z coordinate system and for any point (P) in the system, there is a value of hydraulic head and a value of stream function (explained later). For the vertical and horizontal impermeable boundaries, flux is zero, and may be expressed in terms of hydraulic head gradient as $\left(\frac{\partial h}{\partial x}\right)_{x=0}=0$ and $\left(\frac{\partial h}{\partial z}\right)_{z=0}=0$ respectively, or in

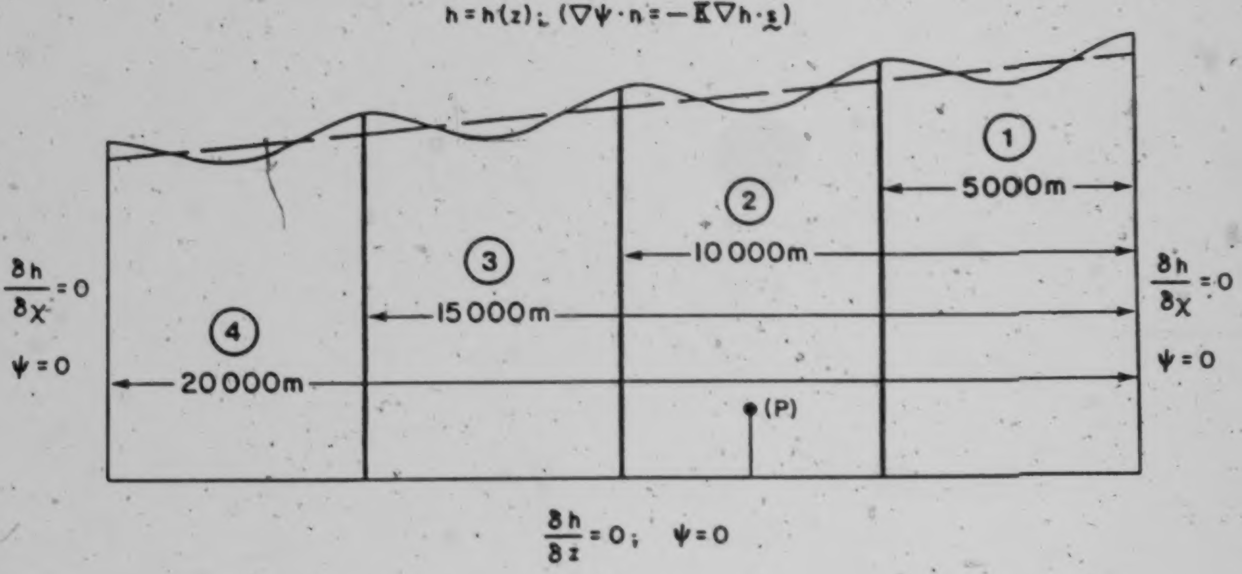

Basin
Sub-basin 2
Incremental Sub-Basins
Sub-basin 1
Slope
DEPTH
WIDTH

$$\frac{\text{DEPTH}}{\text{WIDTH}} = 0.5 \text{ (DEEP BASIN)}$$

$$\frac{\text{DEPTH}}{\text{WIDTH}} = 0.05 \text{ (SHALLOW BASIN)}$$

SLOPE = (3 VALUES USED)

Figure 24. Definition of some basin parameters.


h = h(z); (∇ψ · n = −K∇h · z)
1
2
3
4
5000m
10 000m
15 000m
20 000m
δh/δx = 0
ψ = 0
δh/δx = 0
ψ = 0
(P)
δh/δz = 0; ψ = 0

$$\frac{\delta}{\delta x}\left(\frac{1}{K_z}\frac{\delta\psi}{\delta x}\right)+\frac{\delta}{\delta z}\left(\frac{1}{K_x}\frac{\delta\psi}{\delta z}\right)=0$$

$$\updownarrow \qquad \updownarrow$$

$$\frac{\delta}{\delta x}\left(K_x\frac{\delta h}{\delta x}\right)+\frac{\delta}{\delta z}\left(K_z\frac{\delta h}{\delta z}\right)=0$$

① **Basin number**

Figure 25. Schematic mathematical model of steady-state Plains Regions environment.

all three cases, stream function ( $\Psi$ ) equal zero. Along the water table the pressure is atmospheric and, as a consequence of steady-state conditions, hydraulic head is a function of elevation (z) only, whereas, the stream function may be expressed in terms of the tangential component of the head gradient as (Frind and Matanga, 1985):

$$\nabla\Psi\cdot\eta = -\nabla h\cdot\underset{\sim}{s} \qquad (17)$$

where: $\Psi$ = stream function ($m^3/s$);

$\eta$ = the unit vector normal to the water table (unitless);

$\underset{\sim}{s}$ = the unit tangential vector to the water table (unitless);

h = as previously defined.

The problem involving the Plains Regions environment with boundary conditions as outlined above was formulated in terms of stream functions, which may be defined in two dimensions as (Bear, 1979):

$$\Psi = \Psi \;\; (x,z), \text{ constant along a streamline.} \qquad (18)$$

A streamline on the other hand, is a curve which is everywhere tangential to the specific discharge vector. Thus, streamlines are indicative of the direction of flow at every point in a flow domain. The stream function equivalent of (8), the general partial differential equation that governs the steady-state flow of water in a two-dimensional non-homogeneous medium is (Frind and Matanga, 1985):

$$\frac{\partial}{\partial x}\left(\frac{1}{K_z}\frac{\partial\Psi}{\partial x}\right) + \frac{\partial}{\partial z}\left(\frac{1}{K_x}\frac{\partial\Psi}{\partial z}\right) = 0 \qquad (19)$$

where, every symbol is as defined previously.

With the documentation of (19), a suitable boundary value problem representing regional groundwater flow through the Plains Regions environment has been defined, a numerical solution of which can be obtained quickly and simply using a digital computer.

### 3.3 Method of solution

In the past fifteen years, the finite-element numerical method for approximating solutions to the equation of groundwater flow (Zienkiewicz, 1971; Frind, 1971; Pinder and Frind, 1972; Pinder and Gray, 1977; Wang and Anderson, 1982) has undergone a rapid development for the simulation of complex groundwater flow systems. The model used herein employs the Galerkin finite-element approach (Frind, 1971; Pinder and Frind, 1972; Pinder and Gray, 1977; Frind and Matanga, 1985) which has been well documented over the years. In particular, the computer code developed by Frind (1971) for the solution of potential distribution in two-dimensional cross-sections was used, with slight modifications for the determination of stream function distribution within the region of interest. The model assumes that the flow domain is made up of a continuum of points. This continuum is replaced by a finite number of nodal points in a finite element grid for numerical solution of the boundary value problem. For each nodal point in the field, an integral equation is written to replace the continuous derivatives of (8) or (19). The complete set of all the integral equations for all the nodal points in the field of interest is solved simultaneously. The determined point hydraulic head or stream function values represent the small areas of the field around the points. Details of this method of solution are presented in the papers by Frind (1971), Pinder and Frind (1972), and Frind and Matanga (1985).

In this study, and in the first instance, the computer program (Frind, 1971), which solves (19) for the entire Plains Regions type environment AH (Figure 22) was then executed to obtain the distribution of stream functions. As the original program was written for the solution of potential distribution only, some modifications were necessary. The modifications include: 1. the replacement of Kx by 1/Kz and Kz by 1/Kx during the calculation of the elemental conductance matrix; and 2. the adjustment of the process by which the second-type boundary conditions which now replace the first-type boundary conditions (Frind and Matanga, 1985) of the potential formulation, are read into the program. Furthermore, to facilitate the large data input, and to expedite visual inspection of the results, the computer program was expanded to accommodate a grid generator and a plotting routine. Thus, results were displayed as computer generated flow patterns.

Baseflow was calculated as total groundwater discharge from each sub-basin by integrating between streamlines (Bear, 1979) at the water table in areas where groundwater movement is directed upward. Since baseflow is the fraction of groundwater discharge that occurs as streamflow, the estimated total groundwater discharge will be referred to as "discharge ($q_d$)" in this chapter. It is assumed that baseflow and discharge are directly proportional. It should be mentioned that the use of stream functions herein is valid only because flow is steady, and no sources and sinks are present in the basin.

Occasionally, the hydraulic head distribution in the basin was determined, in order to confirm the resulting flow patterns. This was done by solving (8) with the original computer code of Frind (1971).

To determine possible $q_d$ (A) curves, the effects of basin geometry and geology on the nature of such curves, and hence the nature of regional groundwater discharge through characteristic environments of the Plains Regions, (19) has been solved for the various basin parameters and geologic settings as defined in the previous chapter. Care was exercised to use the estimated characteristic parameters as much as possible, but it was necessary in some instances to expand the range of parameters beyond the estimated values to correct for possibly insufficient available data.

Although the width of the empirically determined type basin was 75 km or greater, its major role was to guide in quantifying the typical hydrogeologic environmental parameters for further analysis. In the theoretical analysis, the width was allowed to vary so that its effects could be analyzed. Moreover, according to Tóth (1963), and as will be seen later in this chapter, it is the [illegible] to width ratio of the basin that matters in groundwater movement rather than e[illegible]meter, separately. In most of the cases simulated, the horizontal distance between the two water divides (s) was 40 000 m (Table 9). The smallest sub-basin had a width (ss) of 5 000 m, and each higher-order sub-basin increased its area upon the preceding one by 5 000 m. In the first instance, eight sub-basins were considered within the main basin. The choice of the size of 5 000 m, and hence eight sub-basins within an area of 40 000 m was not completely arbitrary. It was based on a crude estimate of the ratio of the frequency of occurrence of streams of various orders to specific area sizes in the region of study. Throughout this study, the width of a basin corresponds to its area because a unit width perpendicular to the vertical cross section is assumed.

Two values of 2 000, and 20 000 m were used for the depth to the horizontal impermeable boundary at the lowest valley bottom. These values were selected so that

Table 9. Model variants used in simulations.

Homogeneous cases

| a (m) | n | ns | c′ | ss (m) | s (m) | Figure |
|---|---|---|---|---|---|---|
| 50 | .5 | 8 | .0016 | 5 000 | 40 000 | 27a |
| 50 | .5 | 8 | .02 | 5 000 | 40 000 | 32a |
| 50 | .5 | 8 | .03 | 5 000 | 40 000 | 33a |
| 50 | .05 | 8 | .0016 | 5 000 | 40 000 | 27c |
| 50 | .05 | 8 | .02 | 5 000 | 40 000 | 32b |
| 50 | .05 | 8 | .03 | 5 000 | 40 000 | 33b |
| 50 | .5 | 16 | .0016 | 2 500 | 40 000 | 35b |
| 50 | .05 | 16 | .03 | 2 500 | 40 000 | 35c |
| 50 | .5 | 4 | .0016 | 5 000 | 20 000 | 35a |
| 50 | .5 | 8 | .0016 | 5 000 | 40 000 | 31a |
| 50 | .25 | 8 | .0016 | 10 000 | 80 000 | 31b |
| 50 | .13 | 8 | .0016 | 20 000 | 160 000 | 31c |
| 50 | .07 | 8 | .0016 | 40 000 | 320 000 | 31d |
| 50 | .5 | 8 | .0016 | 40 000 | 320 000 | 31e |

Heterogeneous cases

<table>
<tr><th rowspan="2">equivalent homogeneous case</th><th rowspan="2">Ka/Kt</th><th colspan="3">heterogeneous case figure</th></tr>
<tr><th colspan="2">shallow</th><th>deep</th></tr>
<tr><td>Figure 27a</td><td>100</td><td colspan="2">Figure 37b</td><td>Figure 37a</td></tr>
<tr><td>Figure 27a</td><td>1 000</td><td colspan="2">Figure 37c</td><td></td></tr>
<tr><td>Figure 27c</td><td>100</td><td colspan="2">Figure 38b</td><td>Figure 38a</td></tr>
<tr><td></td><td></td><td>low elevation</td><td>high elevation</td><td></td></tr>
<tr><td>Figure 27a</td><td>100</td><td>Figure 41a</td><td>Figure 41b</td><td></td></tr>
<tr><td>Figure 27a</td><td>1 000</td><td></td><td>Figure 41d</td><td></td></tr>
</table>

a = amplitude of sine curve (representing surface relief)
n = depth/width ratio
ns = number of sub-basins
c′ = regional slope
ss = size of smallest and increamental sub-basins
s = size of main basin

two special cases of the ratio (n) of the depth of the impermeable boundary to the horizontal distance between the two divides would be 0.05, and 0.5. The latter case was selected to represent an extremely deep case, mimicking the thickest probable sequence of sediments in the region. The value of 0.05 on the other hand, represents a shallow case, the environment in which present regional groundwater studies are being conducted, allowing for the generally small widths of basins considered. This value also simulates an instance in which the width of the basin in the deep case is increased greatly with respect to its constant depth.

Three average slopes (c') of the valley flanks, namely, 0.0016, 0.02 and 0.03, and an amplitude of the sine curve (a) of 50 m representing the local relief, were used in all the simulations.

To evaluate the effect of the size of the main basin on the $q_d$ (A) curves, the horizontal distance was reduced to 20 000 m in one simulation, while maintaining the depth to horizontal distance ratio n of 0.5. A slope of 0.0016 was used in this instance. Similarly, the effect of the sizes of the sub-basins on the $q_d$ (A) curves was considered by reducing the smallest basin width and subsequent increments in width from 5 000 to 2 500 m. This condition resulted in sixteen sub-basins, simulating a higher frequency of occurrence of streams of various order within the environment. Two sets of calculations were made, using slope values of 0.0016 and 0.3, for depth to horizontal distance of 0.5 and 0.05 respectively.

The model employed in this analysis requires the knowledge of the ratios of the hydraulic conductivities of hydrogeologic units only. Thus, for the homogeneous case, a hydraulic conductivity value of 1 was used. Heterogeneity was simulated with hydraulic conductivity values of 1 and 100, representing aquitards and aquifers, respectively. Isotropic conditions are assumed in all the simulations in this chapter. As discussed in section 2.3.3.1, the distinction of aquifers from aquitards was based on relative groundwater productivity controlled mainly by hydraulic conductivity. Thus, an aquifer for one use type may become an aquitard for another use purpose. For all the analysis in this thesis, the lower limit of the distinction is arbitrarily set as two orders of magnitude difference in hydraulic conductivity. The upper limit of approximately six orders of magnitude is marked by the ratio of the highest ($5.6 \times 10^{-4}$ m/s) to the lowest ($2.8 \times 10^{-10}$ m/s) hydraulic conductivity (Table 3). In most of the simulations involving the simultaneous existence of aquifers and aquitards, the lower limit was used because it appears to be the most commonly occurring situation, according to available data. In only two cases the aquifer to aquitard hydraulic conductivity ratio was increased to three orders of magnitude. The effects of heterogeneity were analyzed with the ATE and ATR type environments (Figure 22) as the basic units, and the position of the aquifers was varied.

To demonstrate the graphical nature of the results, and to show insight into the method of calculation of discharge while maintaining brevity, only figures which explain the main types of $q_d$ (A) curves and deviations thereof are included in this text. The process of estimation of specific discharge is illustrated in Figure 26. In each sub-basin, the length of the water table in which streamlines are directed outward from the system (discharge area) is noted. The numerical difference between the two stream functions $\psi a$ and $\psi b$ bounding the discharge area is equal to the Darcy discharge according to the equation (Bear, 1979):

$$\Delta Q = \Psi b - \Psi a \qquad (20)$$

where: $\Delta Q$ = Darcy discharge (areal discharge) across the discharge area ($m^3/s$).

To calculate the specific discharge, (20) is divided by the area of the discharge area as

$$q_{d1} = \frac{\Delta Q_1}{A_1} = \frac{\Psi b - \Psi a}{A_1} \qquad (21)$$

where: $q_{d_1}$ = specific discharge through sub-basin 1 ($m^3/s/m^2$);

$A_1$ = area of the discharge area in sub-basin 1 ($m^2$).

### 3.4 Results and their implications

The results of all the calculations in all the cases evaluated, and their corresponding $q_d$ (A) curves are displayed and discussed in the next few paragraphs.

#### 3.4.1 Regularity

One main type of regularity is revealed from an observation of the theoretical $q_d$ (A) curves (Figure 27a). The curve shows an initial high rate of increase in specific discharge with basin area. This high rate is followed by a rate of increase that is much reduced, which in turn, is followed by an intermediate rate of increase that is restricted to the end of the curve at the largest basin area. The flow pattern from which this $q_d$ (A) curve was generated is that of Figure 26. Other basin configuration parameters employed in the calculation of the flow pattern

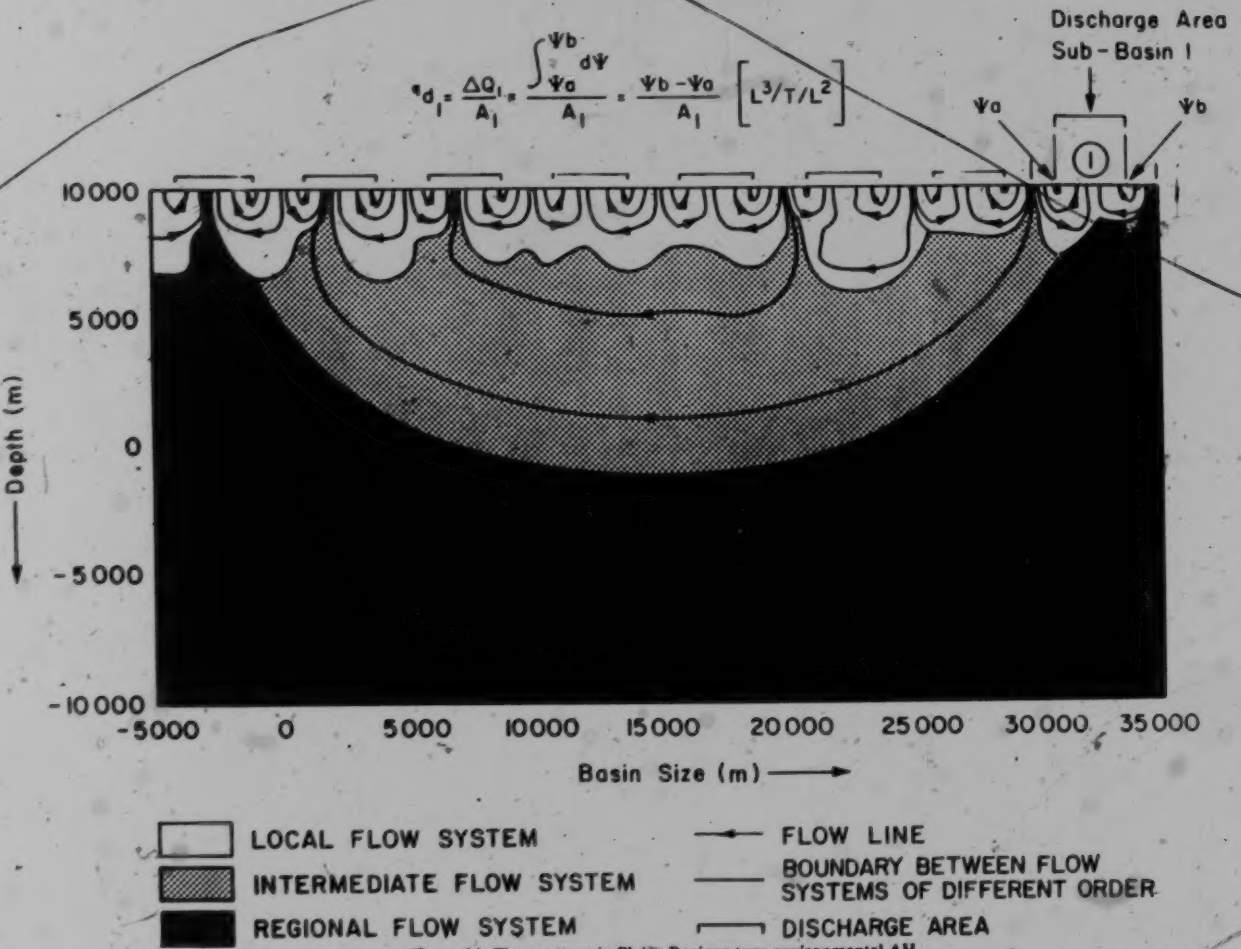

Discharge Area
Sub-Basin I
$q_{d_1} = \frac{\Delta Q_1}{A_1} = \frac{\int_{\Psi a}^{\Psi b} d\Psi}{A_1} = \frac{\Psi b - \Psi a}{A_1} \left[ L^3/T/L^2 \right]$
Ψa
Ψb
1
Depth (m)
10000
5000
0
-5000
-10000
-5000
0
5000
10000
15000
20000
25000
30000
35000
Basin Size (m)
LOCAL FLOW SYSTEM
INTERMEDIATE FLOW SYSTEM
REGIONAL FLOW SYSTEM
FLOW LINE
BOUNDARY BETWEEN FLOW SYSTEMS OF DIFFERENT ORDER
DISCHARGE AREA

Figure 26. Flow pattern in Plains Regions type environmental AH.

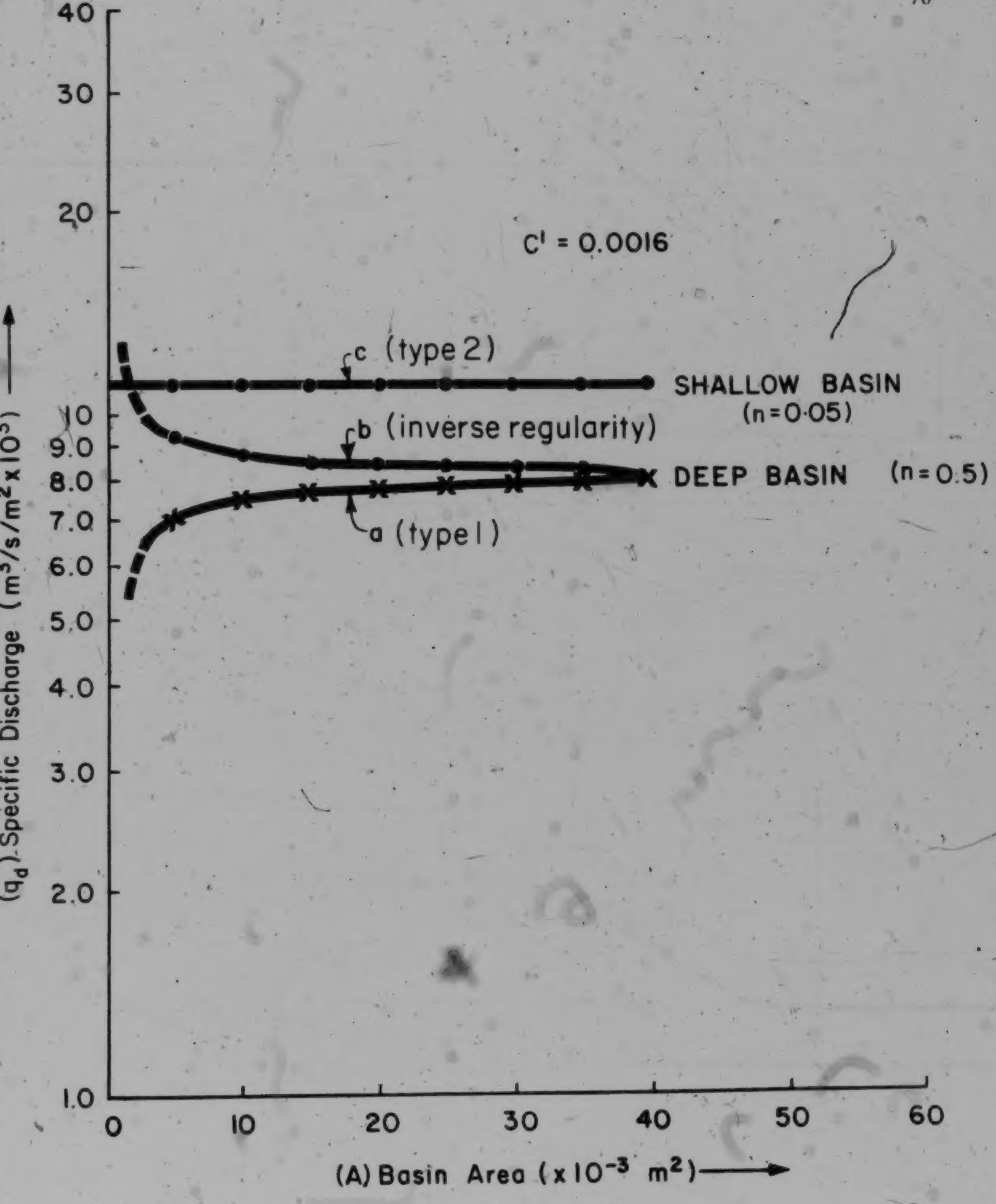


Figure 27. Theoretical $q_d(A)$ curves of AH environment for c' of 0.0016.

are indicated in Figure 28, the corresponding hydraulic head distribution which was determined in order to check the accuracy of the streamlines. Due to scale problems, the actual surface topography cannot be shown in the computer plot, and has been sketched in Figure 28. During the simulations with the finite element model, the water table was assumed to coincide with the surface topography and used in the calculation of the hydraulic head and stream function distrbutions. It should be constantly remembered that such a surface topography is superimposed upon all the figures that show the potential and flow distributions in this chapter.

The $q_d$ (A) curve in Figure 27a has been constructed on the presumption that the smallest sub-basin occurs at the highest topographic elevation, and that areas of sub-basins increase toward lower elevations. The depth of flow and other hydrogeologic environmental parameters in this case are adequate to generate regional and intermediate flow systems in addition to local systems. The generation of these intermediate and regional systems entails the loss of some of water recharged at higher topographic elevations to areas of lower topographic elevations. The transfer of water takes place in a regular manner, such as demonstrated by Figure 27a. It should be noted that the curve has been extended to approach the zero basin area line assymptotically. This is thought to be the case because as the area of the smallest sub-basin at the highest topographic elevation approaches zero, specific discharge through the sub-basin should approach zero slowly. It is noteworthy hence, that the computed discharge pattern is similar in every respect to the previously discussed empirically determined q(A) curve (Section 2.1.2.3). A comparison of Figure 27a and Figures 7, 9, and 12, together with the discussion in the preceding paragraphs indicate that the empirically determined baseflow regularity has been documented theoretically according to the concepts of regional gravity flow of groundwater.

An "inverse regularity" is also displayed in Figure 27b. This regularity type is more difficult to conceptualize because it represents a case of increasing area of sub-basin from low to high topographic elevation (eg, in moving from point J to K, Figure 14). This case is less common in nature. The curve shows conditions of rapidly decreasing discharge with increasing basin area, initially. This is followed by a reduced rate of decrease, and then an intermediate rate of decrease at the scale of the main basin. For an infinitely small sub-basin at the lowest topographic elevation of the setting, specific discharge should be infinely large as indicated by the extension of Figure 27b toward the zero basin area line.

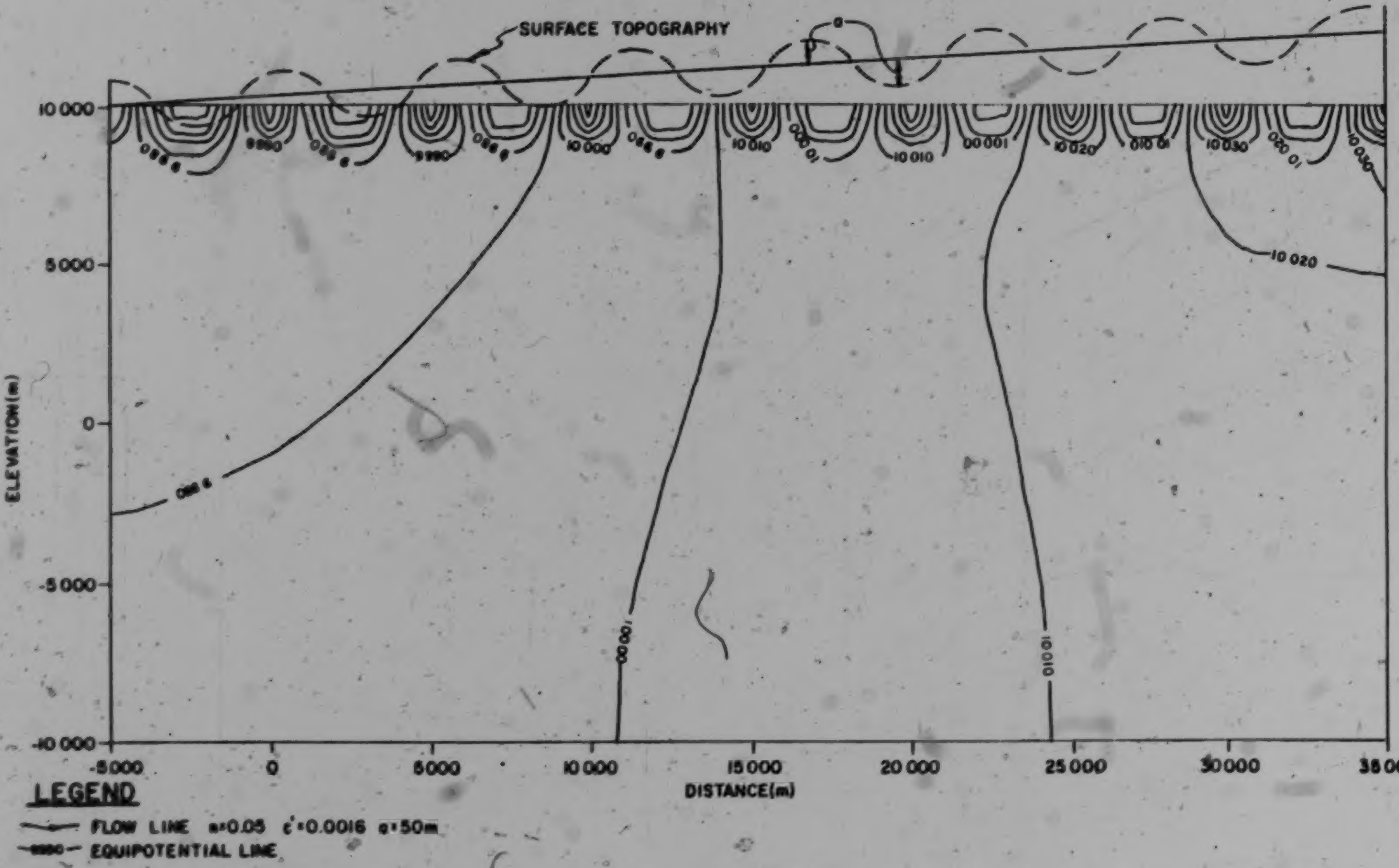


Figure 28. Potential distribution in deep AH type environment.

Thus far, a discharge regularity and its inverse have been obtained that describe the general nature of regional groundwater discharge in the Plains Regions of Alberta. The regularity suggests that specific discharge may be related to the position of the sub-region of flow within the main region in a regular pattern, and that the relation can be explained by our knowledge of the theory of regional gravity flow of groundwater. Various environmental parameters which affect the regional movement of groundwater through porous media could be expected to influence the regularity in a predictable manner. The effects of these parameters will now be investigated. Since the inverse regularity will only seldom occur, attention will be focussed upon the actual regularity.

### 3.4.2 Functional relations

The factors considered include: 1. the ratio (n) of the depth to the horizontal impermeable boundary to the width of the main basin; 2. the average regional slope (c'); 3. the half width of the main basin (s); 4. the width of the smallest sub-basin (ss); and 5. the size and position of aquifers within the system.

The effect of n may be analyzed by comparing Figures 27a and 27c. To produce Figure 27c, an n value of 0.05, obtained by reducing the depth to the impermeable boundary of Figure 26 by a factor of ten while keeping the width of the main basin constant, was used. This reduction is equivalent to increasing the width of the main basin by a factor of ten while maintaining the depth to the impermeable boundary and the eight sub-basins in the environment. In this case, the width of the smallest sub-basin and subsequent increases in the widths of sub-basins become 50 000 m. This configuration is meant to simulate the often shallow environment of groundwater studies in the region of study. The flow pattern from which Figure 27c was constructed is shown in Figure 29. Unlike Figure 27a which shows specific discharge that increases at different rates with increasing basin area, Figure 27c is a straight horizontal line indicating a constant specific discharge for the entire range of basin area. The shallow nature of the environment in this case does not allow for interbasinal flow of water. Thus, each sub-basin has an isolated flow system, unrelated to the adjacent systems. The potential distribution in the setting is presented in Figure 30 with some streamlines to illustrate the self-contained local flow systems.

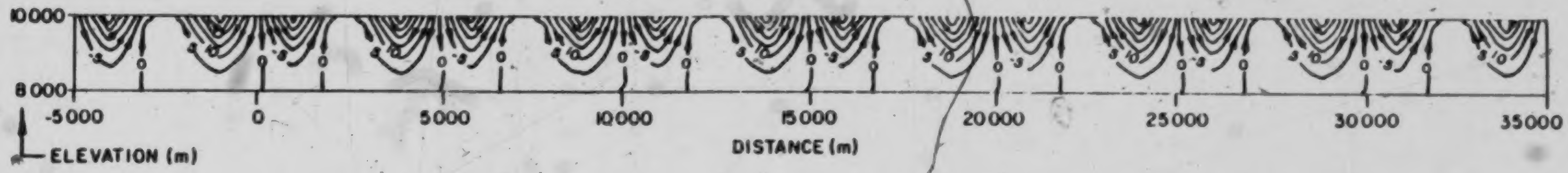


Figure 29. Flow pattern in shallow AH type environment.

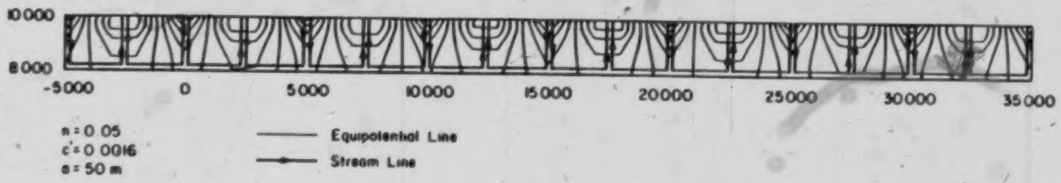


Figure 30. Potential distribution in shallow AH type environment.

For convenience, and to simplify the presentation of the effects of the other parameters, the deep-basin regularity (Figure 27a) and the shallow-basin regularity (Figure 27c) will be referred to as regularity types 1 and 2, respectively.

A more detailed effect of the ratio n was analyzed by increasing the width of the main basin by a factor of two in successive simulations while the depth to the impermeable boundary and number of sub-basins were kept constant for the deep basin. This procedure resulted in successive n ratios of 0.5, 0.25, 0.13 and 0.07 corresponding to smallest and incremental sub-basin widths of 5 000, 10 000, 20 000 and 40 000 m, respectively. The $q_d$ (A) curves corresponding to the ratios simulated are shown in Figure 31. Comparison of Figure 31a, 31b, 31c and 31d shows different forms of regularity type 1 changing gradually into regularity type 2 as n decreases. This situation signifies a gradual change from a flow pattern consisting of local, intermediate and regional flow systems at Figure 31a to a flow pattern of predominantly isolated local systems at Figure 31d. Following these findings, the depth of the system at Figure 31d was increased to produce an n value of 0.5 at Figure 31e. It is evident from a comparison of Figures 31d and 31e that the ratio n is the sole control of the changes in flow pattern rather than the sizes of the smallest and incremental sub-basins. Regularity type 2 is transformed back to regularity type 1 with increase in n.

A comparison of Figures 27c, 32b and 33b indicates that increase in regional slope results in higher specific discharge in basins at lower topographic elevations relative to basins at higher elevations. This statement is also supported by an observation of Figures 27a, 32a and 33a. The ultimate effect is a change in shape of the $q_d$ (A) curves as regional slope increases. In particular, as regularity type 2 changes to regularity type 1 with increase in slope, regularity type 1 is transformed into a linear curve (Figures 32a and 33a) that will be termed as regularity type 3 in this thesis. The flow pattern associated with regularity type 3 is displayed in Figure 34. Figures 34, 32a and 33a indicate that discharge may not occur at all in small basins at high topographic elevations, thereafter, specific discharge increases linearly with increase in area of basin. This situation is comprehensible because, at fairly high regional slope and relatively low local relief, the entire flow region behaves as though it were a unit basin (explained later). Under these conditions, high topographic elevations are recharge areas, low topographic elevations are discharge areas, and the midline acts as a hingeline along which recharge or discharge is zero (Tóth, 1962; Davis, 1963). Figures 32b and 33b also suggest that even in

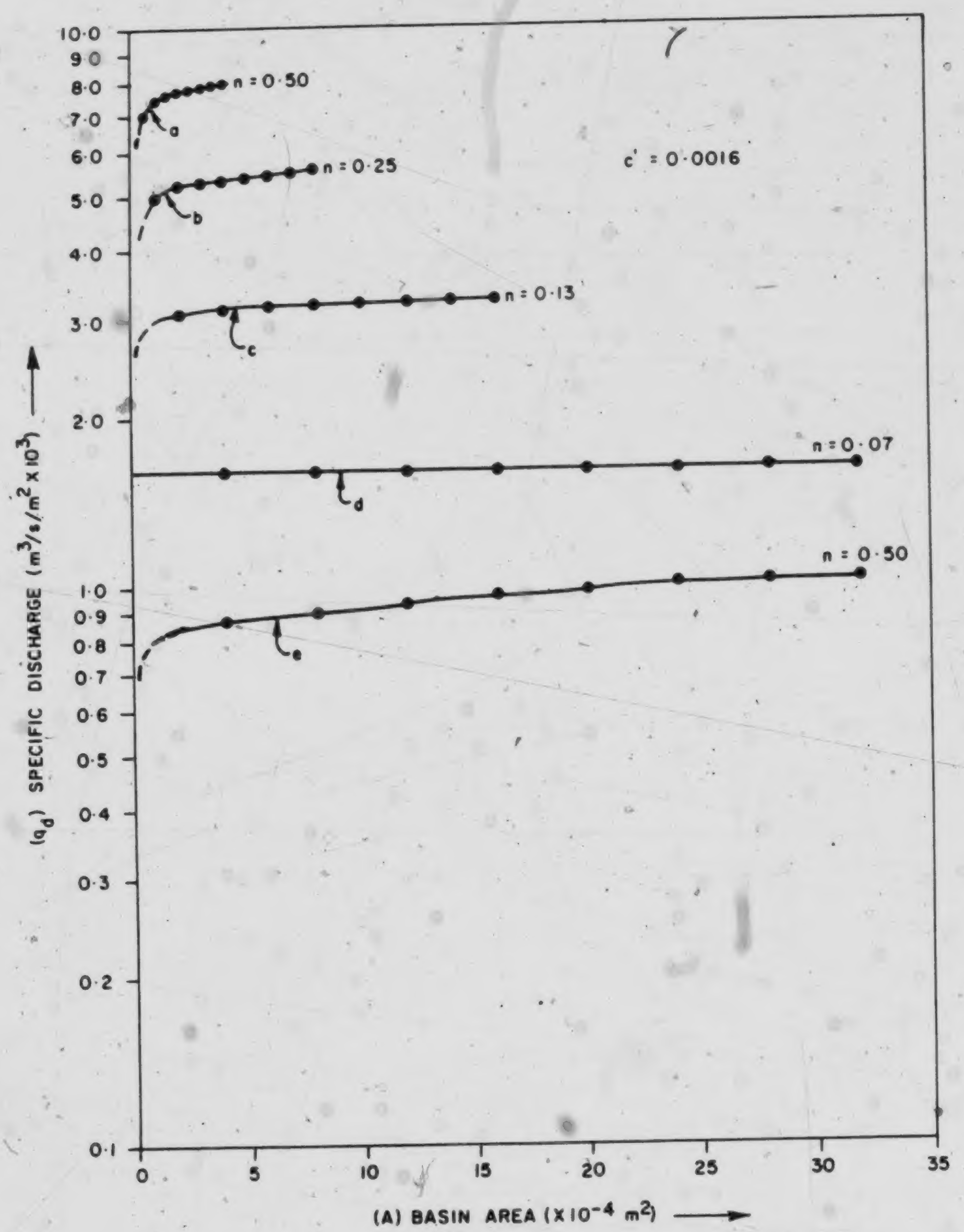


Figure 31. Theoretical $q_d$ (A) curves for varying n values.

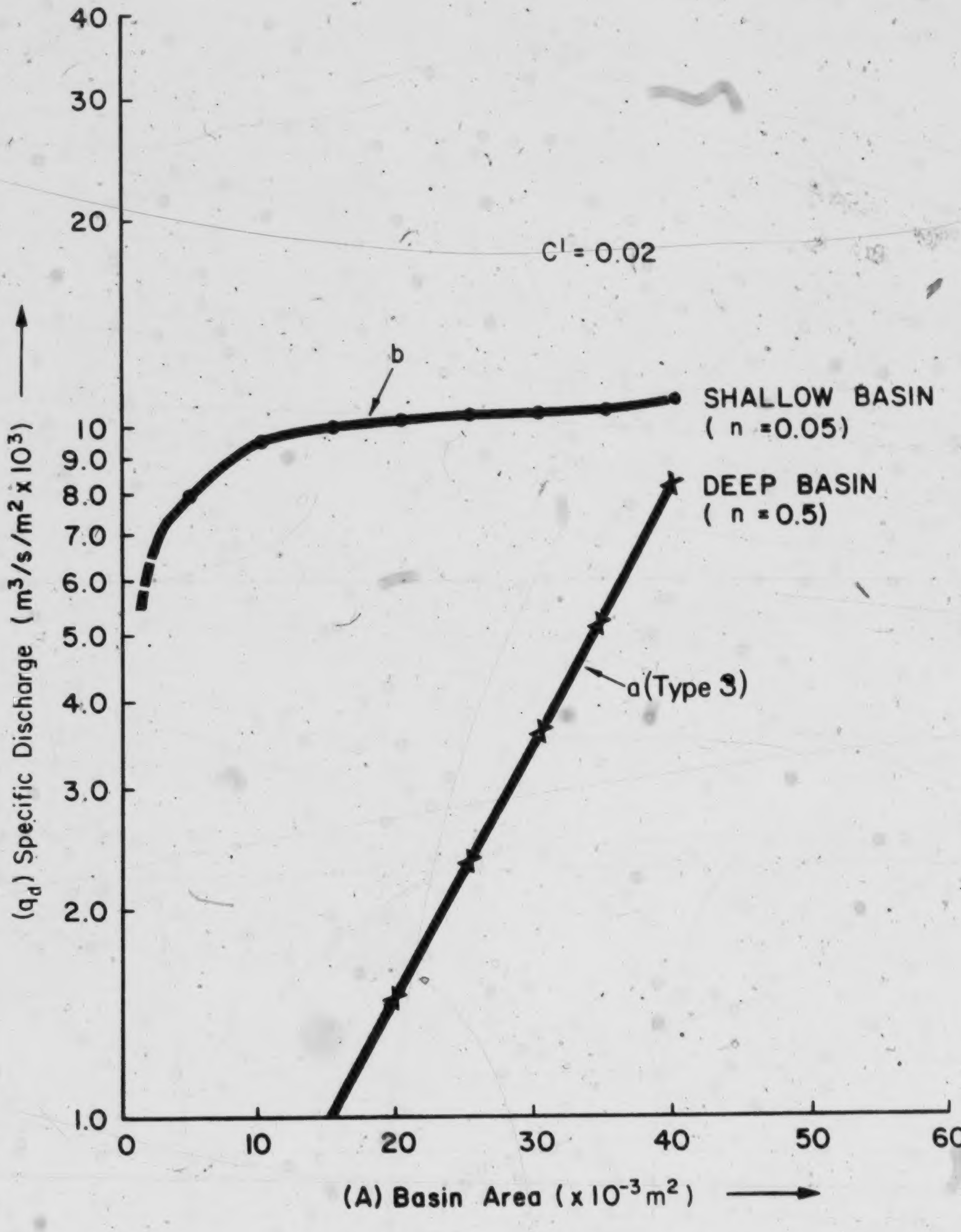


Figure 32. Theoretical $q_d$ (A) curves of AH environment for c' of 0.02.

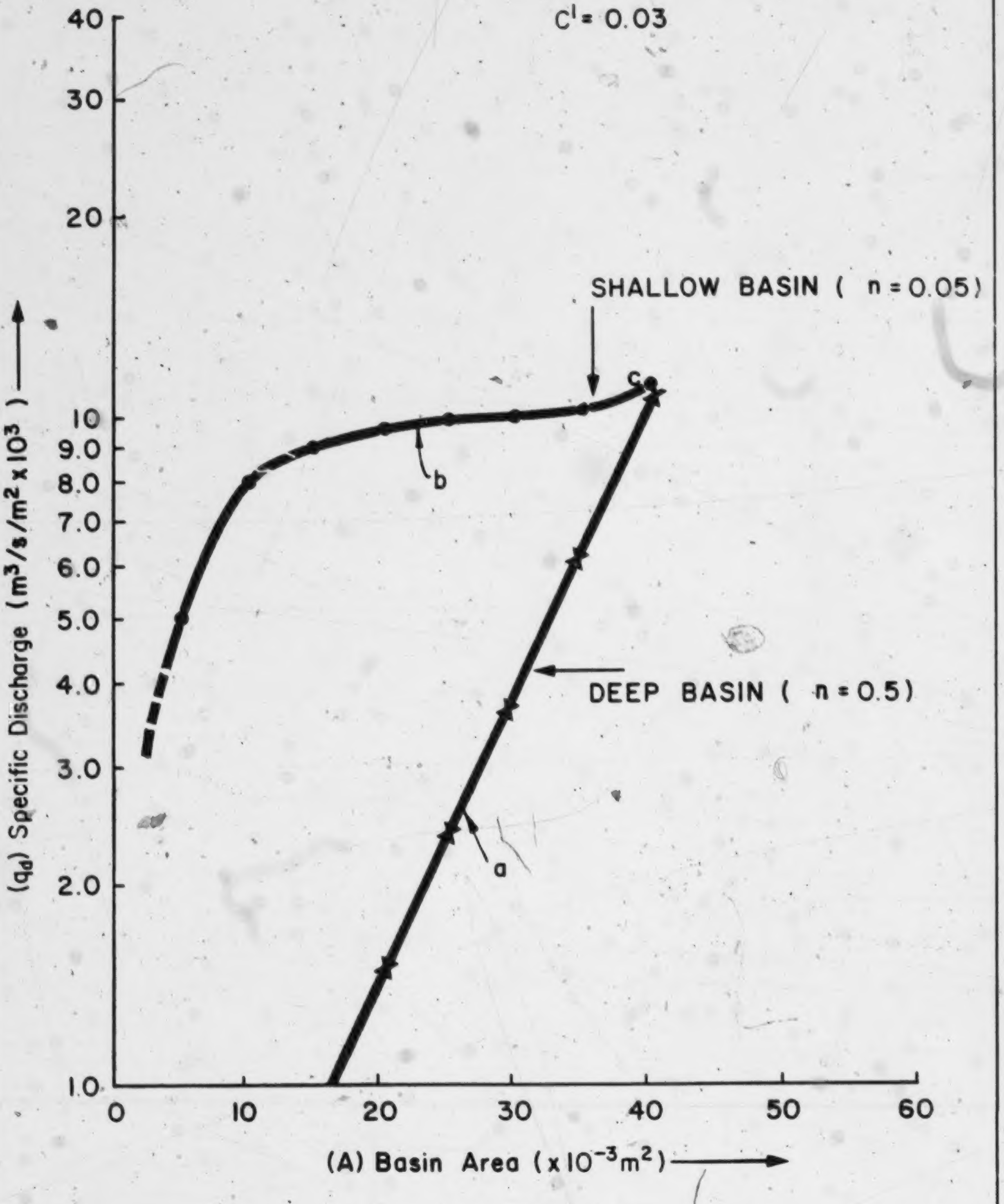


Figure 33. Theoretical $q_d$ (A) curves of AH environment for c' of 0.03.

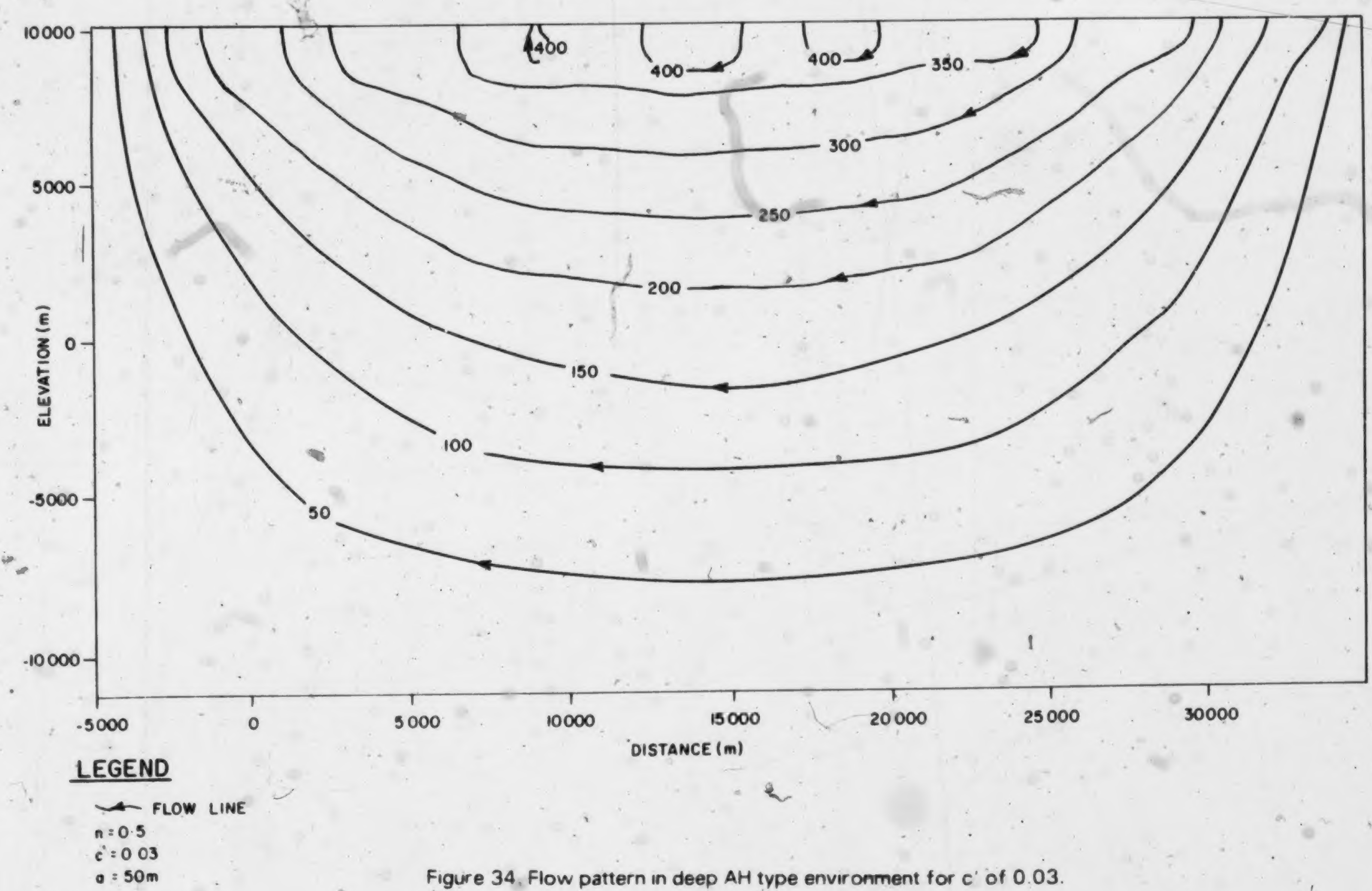


Figure 34. Flow pattern in deep AH type environment for c' of 0.03.

shallow basins, at fairly high regional slopes, such as estimated for some local environments in the Plains Regions, intermediate and/or regional groundwater flow systems might be developed to cause interbasinal movement of water. Recalling that the highest basin area in the discharge plots is synonymous with the lowest topographic elevation in the environment (Section 2.1.2.3), it is clear from Figures 32b and 33b that increase in slope enhances specific discharge at the lowest elevation (point C, Figure 33b) relative to higher elevations. This is an indication that most of the water lost through regional flow systems from the highest elevation ranges is discharged in the lowest areas.

An observation of Figures 27a and 35a explains the effect of the size of the main basin on the $q_d$ (A) curves. To produce Figure 35a, the simulated basin had all the parameters of Figure 26 except that the width and depth were reduced to 20 000 and 10 000 m, respectively. In the presence of depths of basins that are appropriate for the development of regional flow systems, redistribution of water from sub-basins at high elevation to sub-basins at low elevation starts early, even in main basins of small sizes (Figure 35a). However, all three sections of the $q_d$ (A) curve are better developed as the main basin area increases (Figure 27a).

A comparison of Figures 27a, 31e, 35b and 35c shows that the size of the smallest and incremental sub-basins considered, produces no significant change on the nature of the $q_d$ (A) curves. All smallest and incremental sub-basin sizes of 2 500, 5 000 and 40 000 m with a regional slope of 0.0016 produce regularity type 1. For each of these simulations, the dimensions of the environment were adjusted so as to maintain a constant ratio n of 0.5. This is a further evidence that the ratio n, rather than depth or width alone, is a principal controlling factor of the types of flow systems that will form.

The effect of geology was evaluated by analyzing the ATE and ATR type environments (Figure 22) simulating high permeability layers and lenses (aquifers), respectively. The position of the aquifers was varied in the deep and shallow basins. Figure 36 illustrates the positioning of the extensive aquifer during the analysis. The selection of these positions is based on the characteristic occurrence of bedrock and surficial aquifers in the Plains Regions as discussed previously (Section 2.3.3.1). All basin parameters pertinent to this simulation are shown in Figures 36 and 37.

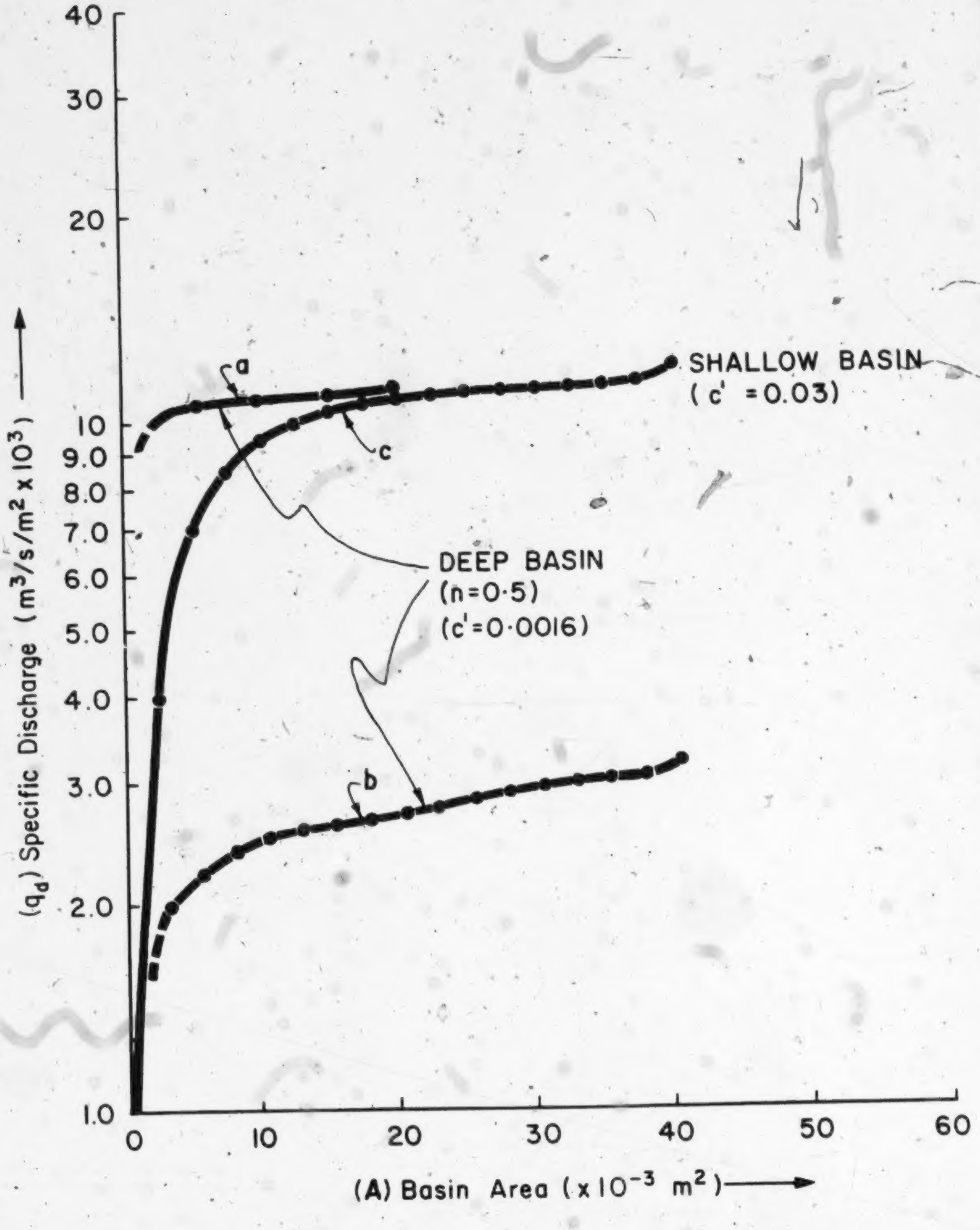


Figure 35. Theoretical $q_d$ (A) curves for narrow and small sub-basins.

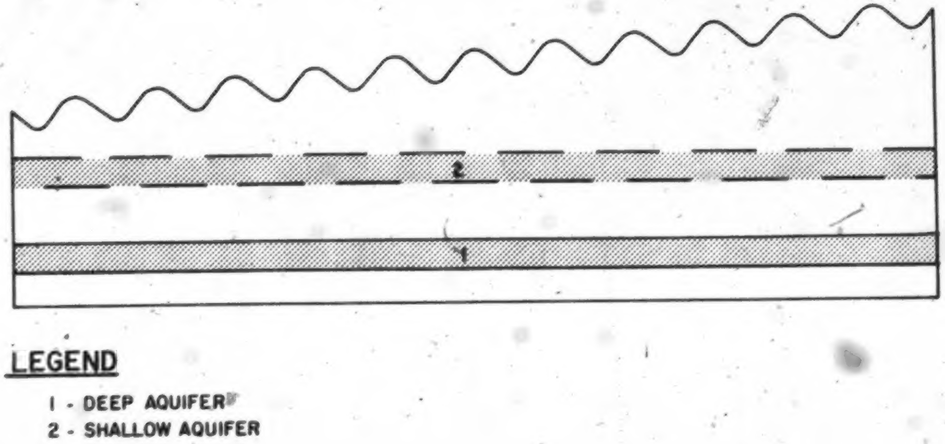

2
1
LEGEND
1 - DEEP AQUIFER
2 - SHALLOW AQUIFER

Figure 36. Illustrations of aquifer positions in ATE environment.

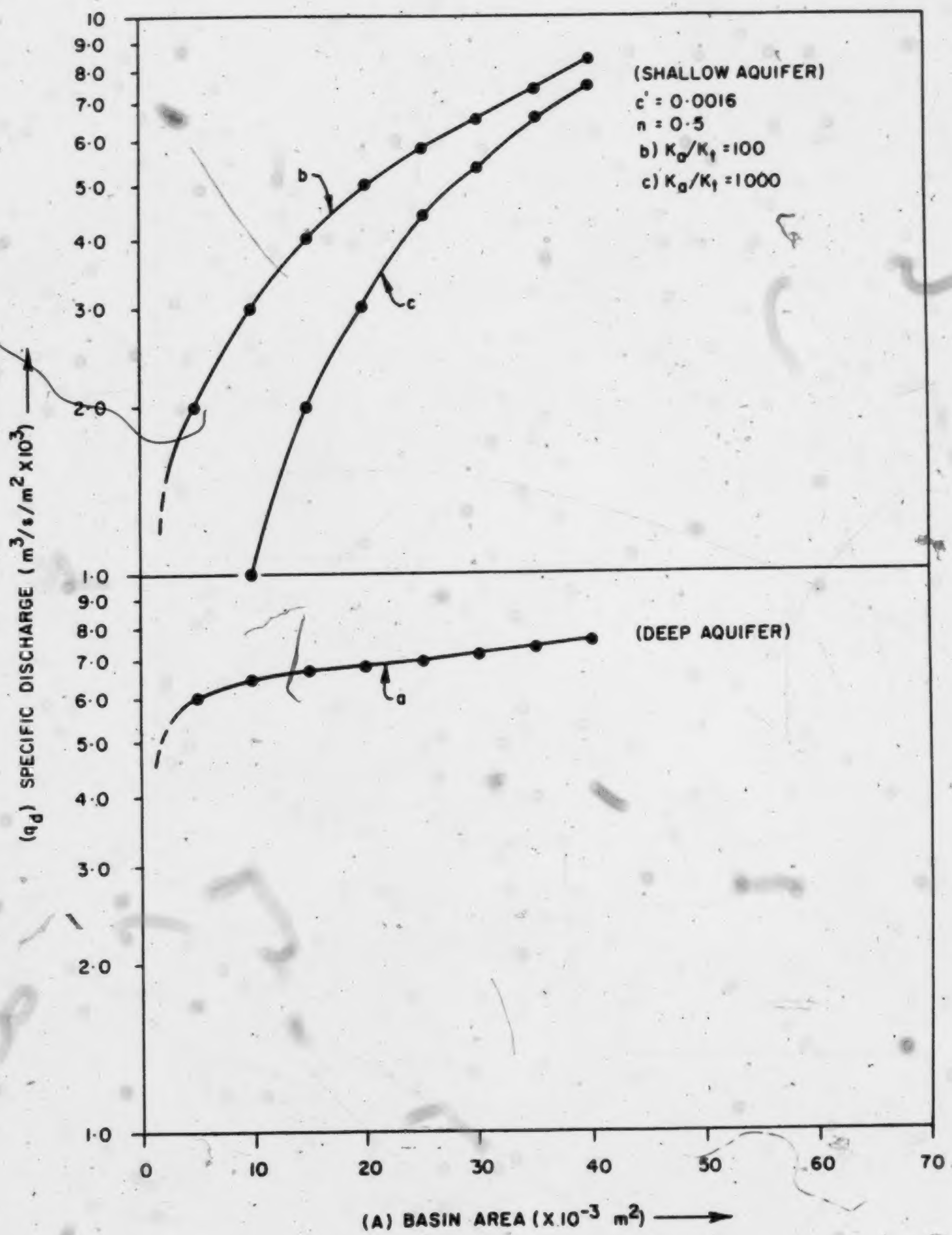


Figure 37. Theoretical $q_d$ (A) curves for deep ATE type environment.

By comparing Figure 27a and 37a, the effect of an areally extensive and deep aquifer in a deep basin is detectable. The aquifer enhances the transfer of water from sub-basins at high elevations to those at lower elevations. In the case of the shallow aquifer in the deep system, by the comparison of Figures 27a and 37b, the enhancement of the transfer of water from high to low sub-basins is so great that the $q_d$ (A) curve (Figure 37b) tends toward linearity. In Figure 37c, the aquifer to aquitard hydraulic conductivity ratio of the preceding case was increased from two (100) to three (1 000) orders of magnitude. There is, again, a further enhancement of the transfer of water, such that discharge in the smallest and topographically highest sub-basin is negligible.

To investigate the effect of the extensive aquifer in the shallow basin, the ratio n of the deep basin (Figure 36) was reduced from 0.5 to 0.05 by reducing the depth to the impermeable boundary by a factor of ten. A comparison of Figures 27c and 38a shows that an extensive aquifer existing even at depths in a shallow environment starts to unite some of the otherwise isolated local flow systems. The interbasinal transfer of water is felt in all sub-basins from high to low topographic elevations. The process of interbasinal water transfer is apparent also, when the aquifer occurs at shallow depths in the shallow environment (Figures 27c and 38b). The process results in the complete transformation of regularity type 2 to regularity type 1. That is, intermediate and/or regional flow systems are now developed in addition to the predominantly local flow systems as evident from a comparison of Figures 29 and 39.

In general, an areally extensive aquifer in an environment acts as a conduit which tends to collect and transmit water from high to low sub-basins. This conclusion is in agreement with the effects of high permeability bodies as predicted by Tóth (1962) and Freeze and Witherspoon (1967). The effects are expressed quantitatively by the $q_d$ (A) curves.

To investigate the effects of high permeability lenses (partial aquifers) on the $q_d$ (A) curves, the ATR type environment was used. Only the deep basin was considered with the aquifer occurring at a shallow depth, and occupying a low or high topographic elevation (Figure 40) in each of the two simulations. By comparing Figures 27a, 41a and 41b, it is obvious that the presence of a high permeability lens tends to give regularity type 1 a more linear shape, signifying a more linear distribution of specific discharge among sub-basins of an environment. In general, groundwater discharge in sub-basins at lower elevations is enriched at the expense of discharge in high sub-basins. Also, discharge in sub-basins at the up-slope end

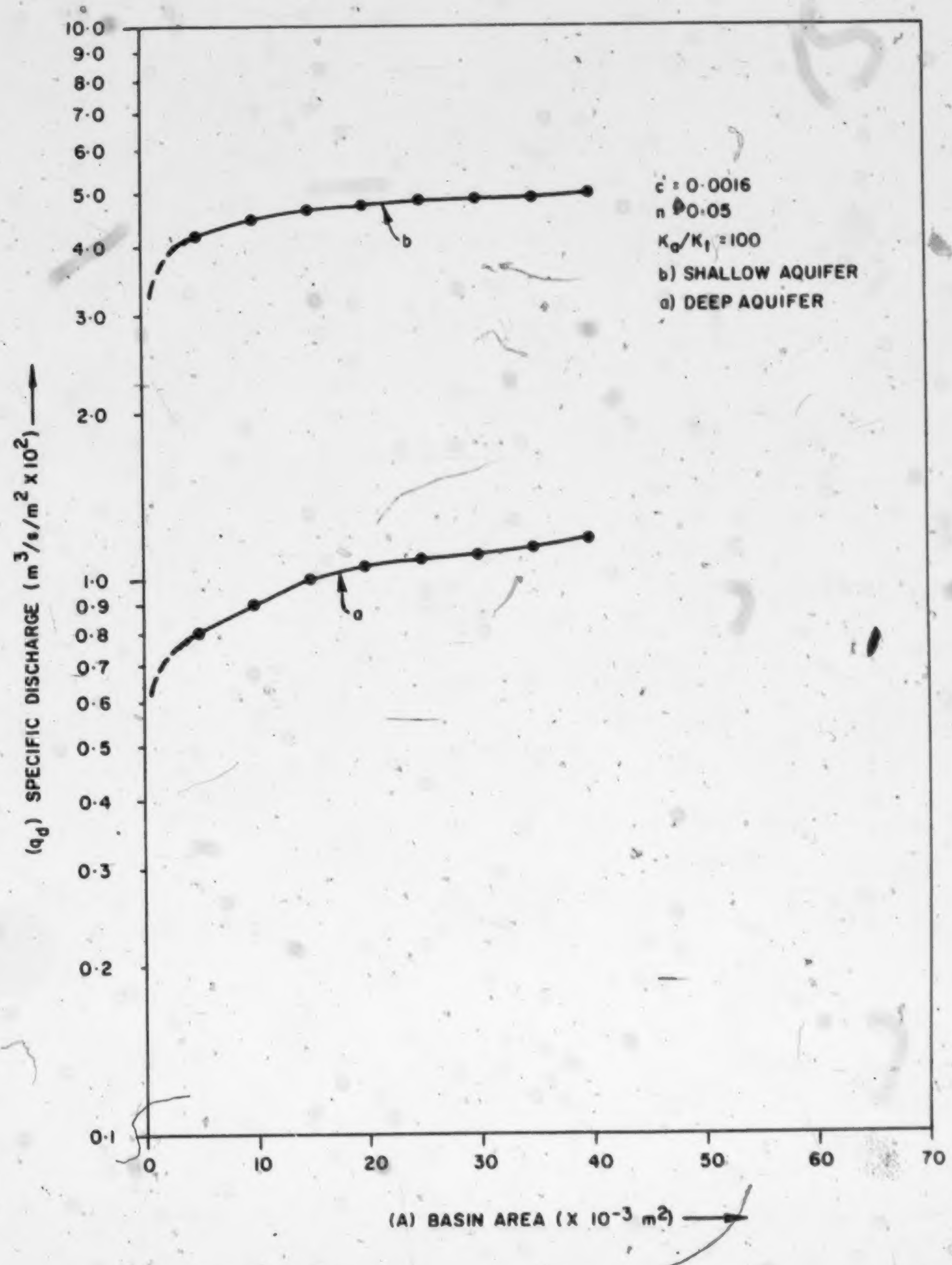


Figure 38. Theoretical $q_d$ (A) curves for shallow ATE type environment.

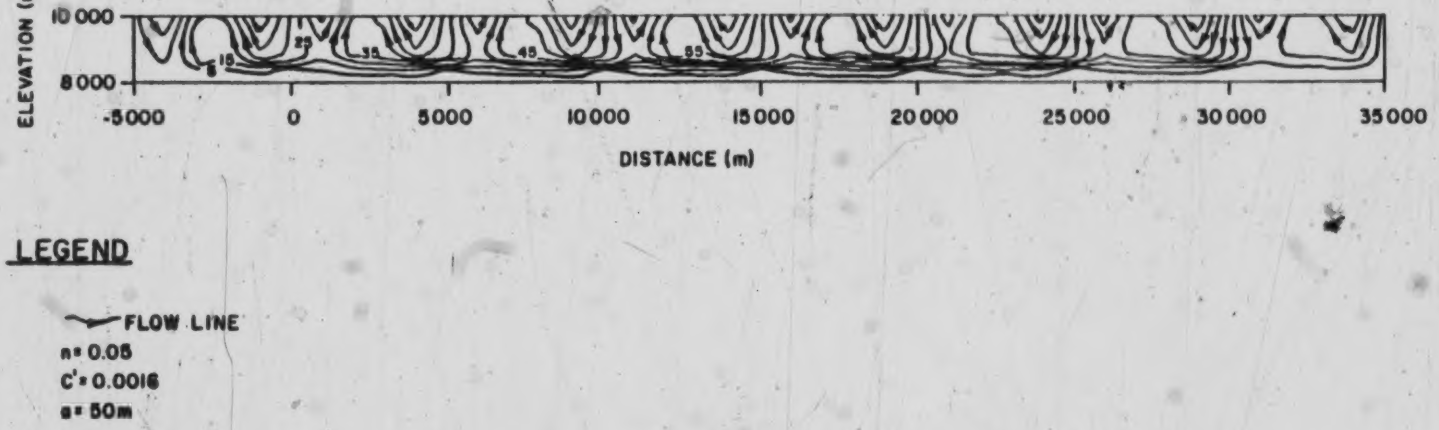


Figure 39. Flow pattern in shallow ATE type environment with deep aquifer.

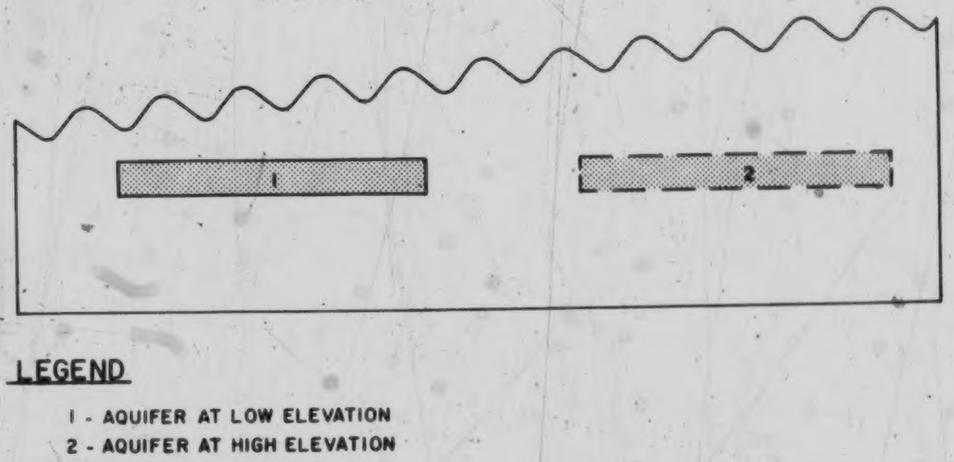

1
2
LEGEND
1 - AQUIFER AT LOW ELEVATION
2 - AQUIFER AT HIGH ELEVATION

Figure 40. Illustration of aquifer positions in ATR environment.

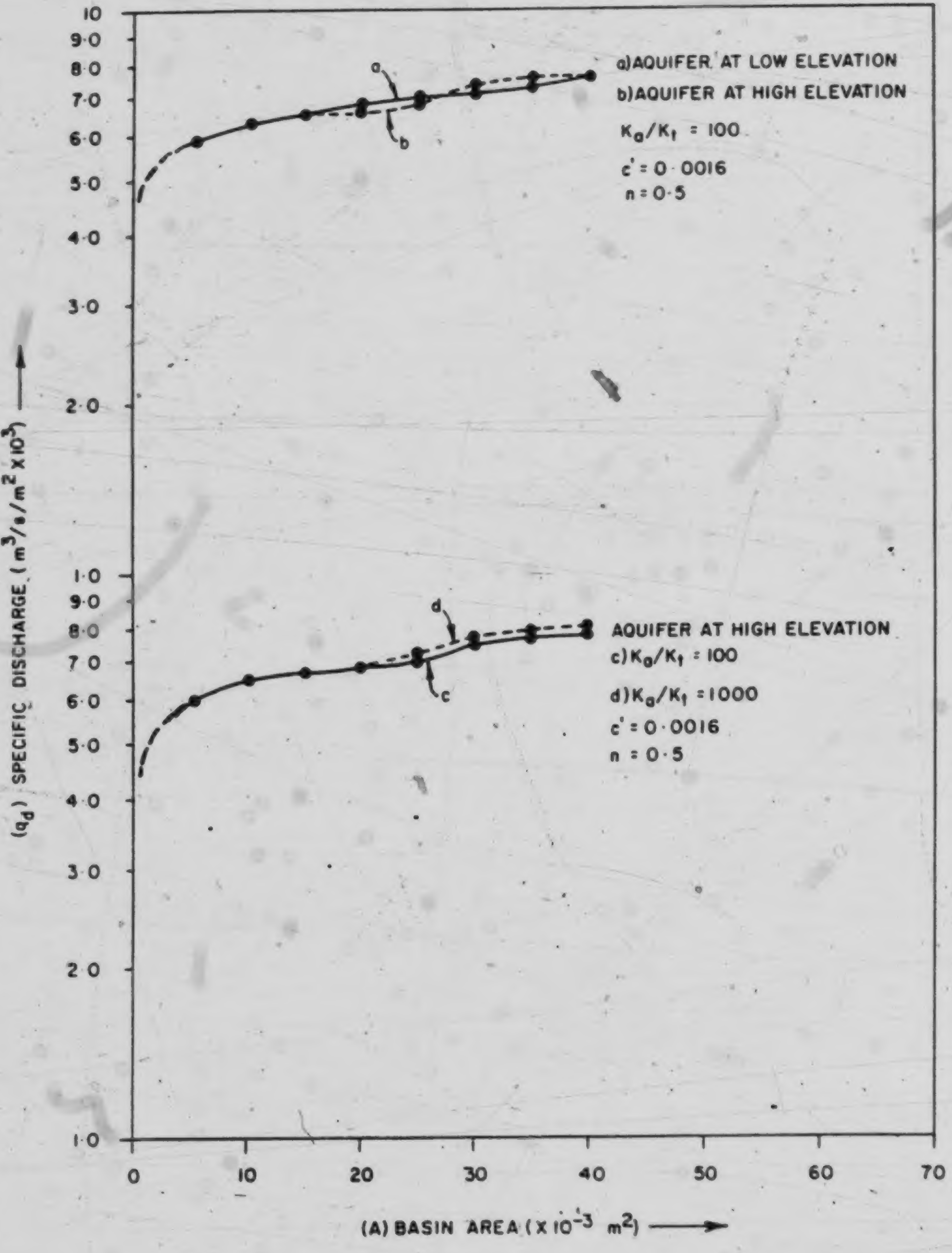


Figure 41. Theoretical $q_d$ (A) curves for ATR type environment.

of the lens is preferentially depleted relative to that of the sub-basins down-slope of the lens (Figure 41a and 41b). In Figure 41d, the aquifer to aquitard hydraulic conductivity ratio was increased to three orders of magnitude (1 000) relative to the two orders of magnitude (100) of Figure 41c. Such a change in hydraulic conductivity results in a minor enhancement of the effects of the lenses as discussed above.

The effects of variation within the range of hydraulic conductivities expected for geologic materials in the Plains Regions is such that no new $q_d(A)$ curves are developed, other than those for the homogeneous cases. The regular pattern of regional discharge of groundwater is not destroyed by the occurrence of high permeability layers and lenses. Only the conversion of one type of regularity to the other is possible under certain conditions, for instance, the modification of regularity type 2 to regularity type 1 (Figures 27c and 38b) by the presence of a shallow high permeability layer in a shallow environment. On the basis of the discharge regularity types therefore, it may be concluded that typical basins in the area of study act hydrologically as to mask the effects of inhomogeneity to a low level. A similar conclusion was reached by Farvolden (1963) in his evaluation of the geologic controls on groundwater storage and baseflow in northern Nevada.

### 3.5 Summary and possible role of discharge regularities

Through the solution of boundary value problems comprising the equation, domain and boundary conditions for various settings in the Plains Regions, it has been possible to approximate an empirically determined baseflow regularity and its related types.

One main discharge regularity (type 1, Figure 42a) which changes into two other types (ie, 2, Figure 42c and 3, Figure 42d) and their corresponding inverse regularities (Figures 42b, 42c and 42e, respectively) were obtained in the present analysis. These regularities indicate that regional groundwater discharge through typical basins of the Plains Regions occurs in a regular manner.

The main regularity type shows specific discharge increasing rapidly with initial increase in basin area, followed by a reduced rate of increase at larger basin areas. This suggests that every other controlling parameter being equal (if ever), specific discharge might be related to the area of a basin of interest.

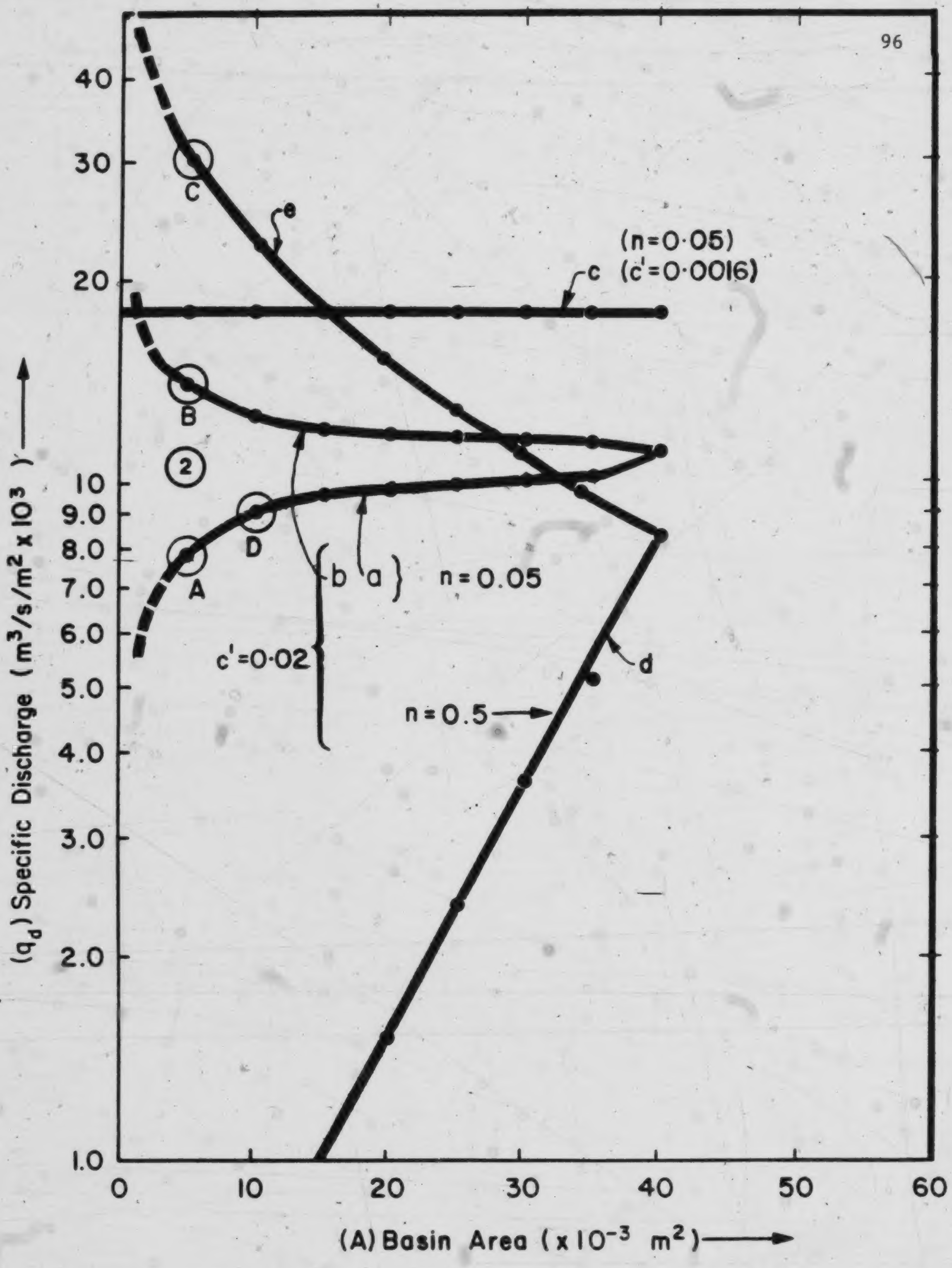


Figure 42. Summary of theoretical $q_d$ (A) curves.

Through numerical analysis, it is possible to show that each regularity type is a reflection of the types of groundwater flow systems which exist in a particular environment.

Analysis of the depths and widths of type environments indicates that the depth to width ratio of the environment is a major control of the type of flow pattern and hence the regularity type that will form. With adequate depth to width ratios, local, intermediate and regional flow systems develop, causing the transfer of water from topographically high to low sub-basins, and forming regularity type 1. At a greatly reduced ratio, depth requirements are not met, therefore, only isolated local flow systems are developed. Under these conditions, interbasinal transfer of water ceases, resulting in a constant specific discharge for all areas of basins which is recorded by regularity type 2.

The effect of increasing average regional slope is to convert a flow pattern of predominantly local flow systems into a flow pattern of local, intermediate and regional flow systems, resulting in the conversion of regularity type 2 to regularity type 1, and then to regularity type 3.

The effect of high permeability bodies is the transfer of more water from high to low sub-basins than would occur in their absence. When these bodies are large enough, they are capable of transforming regularity type 2 to regularity type 1, and probably to regularity type 3. Like increasing regional slope, this effect causes the concentration of regional groundwater discharge in low topographic areas. Small bodies (lenses) modify the shape of the existing regularity without changing it to a completely different type.

The discharge regularities developed in this chapter might be usefully employed in groundwater studies in several ways. Before discussing their role, it must be stated that the theoretical analysis of discharge is equivalent to the determination of the dynamic reserves of groundwater. In particular, the estimated specific basin yield is synonymous with the natural basin yield per unit area of basin (specific natural basin yield) as defined by Freeze (1968), and Freeze and Witherspoon (1968). According to this definition, the specific discharge is the amount of water that can be developed from a unit surficial area of an undeveloped groundwater basin. The regularities may therefore, be employed as follows: First, a knowledge of the discharge regularities in an environment implies a knowledge of the prevailing environmental parameters, and hence the possible flow pattern in the environment. Conversely, a knowledge of the environmental parameters may be used to infer the possible discharge regularity, and then the possible flow pattern. Second, a knowledge of the functional relations between the $q_d$ (A) curves and basin parameters may give guidance to exploitation strategies. For example, according to Figure 42c (regularity type 2, included schematically in this diagram), the same specific basin yield can be developed from all sub-basins and the main basin in that environment. In Figure 42a, a specific basin yield of $9.0\times10^{-3}$ $m^3/s/m^2$ can be developed from the sub-basin of 10 000 $m^2$ in area. Above and below this basin area, sub-basins will produce specific basin yields that exceed $9.0\times10^{-3}$ $m^3/s/m^2$ in small amounts, or that are significantly less than this amount, respectively. Figure 42d shows linearly increasing specific basin yield with basin area to a maximum of $8.25\times10^{-3}$ $m^3/s/m^2$ for the main basin. The quantitative information above might guide at the planning stage of regional exploitation of groundwater. Third, baseflow regularities may also guide in the definition of type environments for regional groundwater studies. Regularity type 2 of a constant specific basin yield represents a case of all sub-basins being type environments for that region. Figure 42a (regularity type 1) reveals a break-of-slope (point D) at a basin area of about 10 000 $m^2$. Relatively speaking, it appears as if conditions of groundwater flow are uniform in sub-basins of at least this area or more. It is possible therefore, to adopt any of such sub-basins as typically representative of groundwater conditions in such regional settings. On the other hand, Figure 42d (regularity type 3) represents a case where the definition of a type basin might be impossible because specific basin yield differs greatly for all sub-basins, and attains a maximum at the scale of the main basin only. Fourth, a significant relationship expressed by the $q_d$ (A) curves, and the relationship that will be explored mainly in the next chapter is that between topographic elevation and specific basin yield. To explain this relationship, it is necessary to remember that the actual regularity types (Figures 42a and 42d) represent cases in which small basin occur at high elevations while larger basins occupy low elevations. On the other hand, the reverse regularity types (Figures 42b and 42e) characterize small basins at low elevations and larger basins at high elevations. Consequently, it is easily recognizable that specific basin yield is greater at low elevations relative to higher elevations. For instance, a basin of 5 000 $m^2$ at a high elevation in a shallow regional setting sloping at 0.02, discharges naturally at $7.8\times10^{-3}$ $m^3/s/m^2$ (point A) whereas the same basin at a low elevation would discharge at $1.4\times10^{-2}$ $m^3/s/m^2$ (point B). Similarly, such a basin when located at a high elevation in a deep regional setting discharges at a negligible rate (omitted from the low part of Figure 42d), whereas, at a low elevation, its discharge would be as high as 3.0x10$^{-2}$ m$^3$/s/m$^2$ (point C). Assuming a unit basin (explained later) that corresponds to the complex regional setting discussed in this chapter, low and high topographic elevations will be equivalent to discharge and recharge areas (Tóth, 1962), respectively. Under these circumstances, the discharge area possesses the excess water that has been transported from the recharge area according to the hypothesis of section 1.3 as illustrated in Figure 4: the $q_d$ (A) curves portray quantitatively the natural regional groundwater hydrology of drainage basins in terms of the functional relation between discharge and size, and between discharge and position within the basin.

It is now clear, and can be quantified that under virgin conditions, more water moves through the discharge areas than the recharge areas. Relative to the average water balance of the setting, excess and deficit of water should be associated with discharge and recharge areas, respectively, as suggested by the $q_d$ (A) curves. It is also known, and is a topic of on-going research that the steady-state distribution of groundwater is modified by development (Theis, 1940; Bredehoeft and Young, 1970; Freeze, 1971; Bredehoeft et al, 1982). In the next chapter, a new approach to the solution of the behaviour of a basin under development is carried out for the purpose of optimal basin development and management, based on the information from the $q_d$ (A) curves.

# 4. RESPONSE OF PLAINS REGIONS GROUNDWATER REGIME TO DEVELOPMENT

## 4.1 General statement

Through a theoretical duplication of the empirically determined baseflow regularity, it has been found that the movement of groundwater through a Plains Regions environment is unevenly but systematically distributed in areas of different sizes. Relative to the recharge area, more water appears to occur in the discharge area under steady-state flow conditions. The curve of the specific discharge is a quantitative expression of this difference. Variations in the parameters of the hydrogeologic environment results in variations in the $q_d$ (A) curves, marking variations in the relative distribution of the quantity of water.

That the quantity of water though flow varies spatially across the environment implies that its exploitation must be planned in order to use it optimally. Exploitation of the water implies its extraction like any other mineral, through pumping wells. The development of an environment through pumping wells changes the recharge-discharge regime of the environment (Bredehoeft and Young, 1970; Freeze, 1971; Bredehoeft et al., 1982), thus, it could be expected to change the pre-existing steady-state distribution of water within the environment. The positioning of the well can therefore be expected to play a major role in determining the ultimate recharge-discharge regime that will result in conjunction with the nature of the existing parameters of the hydrogeologic environment. Finally, the use for which water is intended is a principal determinant of the quantity of water to be considered as optimal.

In this chapter, Plains Regions environment types are developed systematically in order to propose guidelines for regional development and optimal management practices. Well location for optimal basin infiltration from precipitation and optimal basin utilizable water (explained later), and basin stability are of main concern. To be more specific, the problem that is addressed in this chapter may be put in the form of a question as: Does the location of an extraction site within the Plains Regions environment matter from the point of view of optimum utilization of the basin's groundwater resource? If it does, how can this knowledge be included in planning the optimal development and management of groundwater in the regional sense?

## 4.2 Formulation of problem

### 4.2.1 Previous formulations

Although some studies have been conducted toward the solution of the questions above or related types, no one study has analyzed the problem in the degree of detail considered in this thesis. Some previous studies will be reviewed briefly to highlight their differences and similarities to the present study.

In the past few decades research on the response of large groundwater systems to development has concentrated on one issue; that of dispelling the common idea held by most hydrologists that the water balance of an area determines the magnitude of possible groundwater development. Most of the research effort has been expended on the determination and explanation of the adjustments in natural recharge and discharge rates (capture) due to pumping. Although this consideration has often led to indirect conclusions regarding the role of some hydrogeologic environmental parameters in basin management, such conclusions are still incomplete. The effects of the recharge and discharge flow regimes of regional groundwater systems have not been included in any of these studies in detail. Furthermore, restrictions in the specification of the top (water-table) boundary conditions have led to the lack of a detailed analysis of the nature of capture from precipitation, the major source of groundwater recharge, in some numerical modeling studies of groundwater systems. It would appear as if it is for this reason that most previous studies favour the capture of natural discharge relative to the capture of natural recharge into a groundwater system under development. Since capture from precipitation is a major source of input of water into the groundwater system, its consideration is important in a study of the response of a groundwater system to development. Therefore, a method must be sought for its inclusion in the present analysis.

Bredehoeft and Young (1970) determined the effects of development on natural discharge by phreatophytes. Although the location of the withdrawal site was varied, recharge into the system was constant, hence its capture was not discussed. They concluded however, that the response of groundwater systems depends upon the aquifer parameters (transmissivity and storage coefficient), the boundary conditions, and the positioning of the development within the system. Similarly, Young and Bredehoeft (1972) analyzed the effects of pumping on return flows and downstream water availability in a regional system with stream-connected aquifers. Again, capture from precipitation was not discussed as this was not the objective of their study. In a related analysis, Bredehoeft et al. (1982) considered the capture from both natural discharge and recharge. Recharge and discharge were restricted to portions of the system, hence the effects of the capture from precipitation through the recharge and discharge regimes of the system were not analyzed. In his one-dimensional analysis of the recharge mechanism designed to mimic a regional groundwater flow system, Freeze (1969c) concluded that owing to the influence of the variations in groundwater recharge and discharge, infiltration and evaporation rates will show variations even under homogeneous meteorological conditions. However, he did not consider the effect of withdrawal in his analysis.

The role of basin stability in the analysis of the response of a groundwater basin to development is examined in the paper by Freeze (1971). Using a schematic diagram, he illustrated the relationships between the rate of withdrawal, recharge and discharge, and the behaviour of the water table. The rate of withdrawal corresponding to the lowest position of a steady water table is the "maximum stable basin yield" above which, instability in the basin is marked by a rapid drop in water table; that is, groundwater mining results. Freeze did not discuss the effect of the positioning of the withdrawal site in his analysis. In a similar consideration, Kovács (1981) used the maximum point of a characteristic curve of groundwater balance to divide the range of the possible water table into two zones. In the zone above the maximum point, development of the water resources in the basin results in a new equilibrium. If, however, the original water table lies in the lower zone (below the maximum point) groundwater mining results from any amount of withdrawal from the system. Variation of withdrawal site and the recharge and discharge regimes of the system were not considered by Kovács (1981). The establishment of equilibrium conditions during regional groundwater exploitation is an important consideration because one of the major objectives of groundwater management is to ensure that sufficient water is left in subsurface storage for future generations (Chun et al., 1964; Fowler, 1964).

Other formulations of relevant hydraulic groundwater management problems have taken the form of classical mathematical programming in systems optimization (eg. Aguado, 1974; Aguado and Remson, 1974; Schwarz, 1976; Haimes, 1977). In these studies, an optimal solution is obtained in one step, therefore, no room is given for the detailed discussion of the results in a manner that is applicable to basin management. Moreover, the natural steady-state regimes of flow (discharge, midline, recharge) are not considered in most of these studies.

This review shows that all the components of regional groundwater flow have not been considered by any one previous study. These components include: 1. a consideration of the natural steady-state discharge of water and its associated flow regimes; 2. the effects of withdrawal and its systematic variation across the regional setting; 3. the spatial variation in the quantity of water that can be developed from the region; 4. the capture of natural discharge, and recharge from precipitation into the system; 5. the stability of the groundwater basin under conditions of development, and 6. the effects of changes in the parameters of the hydrogeologic environment on the response of the system to development. It is only when these components are analyzed systematically and documented in a logical manner that progress toward their easy pre-consideration in regional groundwater development and management can be made. Some of the aspects are considered in the present formulation.

### 4.2.2 Present formulation

As with the formulation in section 3.2, the shape of the region, existing boundary conditions, and the partial differential equation describing the flow in the region must be defined.

Two-dimensional cross sections in the x-z directions were used with the y-dimension being infinite. The BH category (Figure 22) of Plains Regions environments with some modifications was the basic unit of analysis. The region is bounded at the surface by an axially symmetrical topographic depression with the land surface rising linearly from the valley bottom to the water divides (Figure 43). A horizontal impermeable stratum lies at the base, and vertical impermeable boundaries occur on either side of the region at equal distances away from the valley and divide. The water table is assumed to lie parallel to the land surface. The selection of the vertical impermeable boundaries was not arbitrary. They were located far enough from the sites of development so that no flow conditions could be assumed without influencing the response of the system to development during the transient runs. Although the boundaries were to be located far enough, the further they were, the larger was the domain, and consequently, the more expensive the runs would be. The anticipated large number of runs required that the boundaries be located optimally. For this purpose, several model runs were executed during which the positions of the boundaries were varied. First, the right boundary was located along

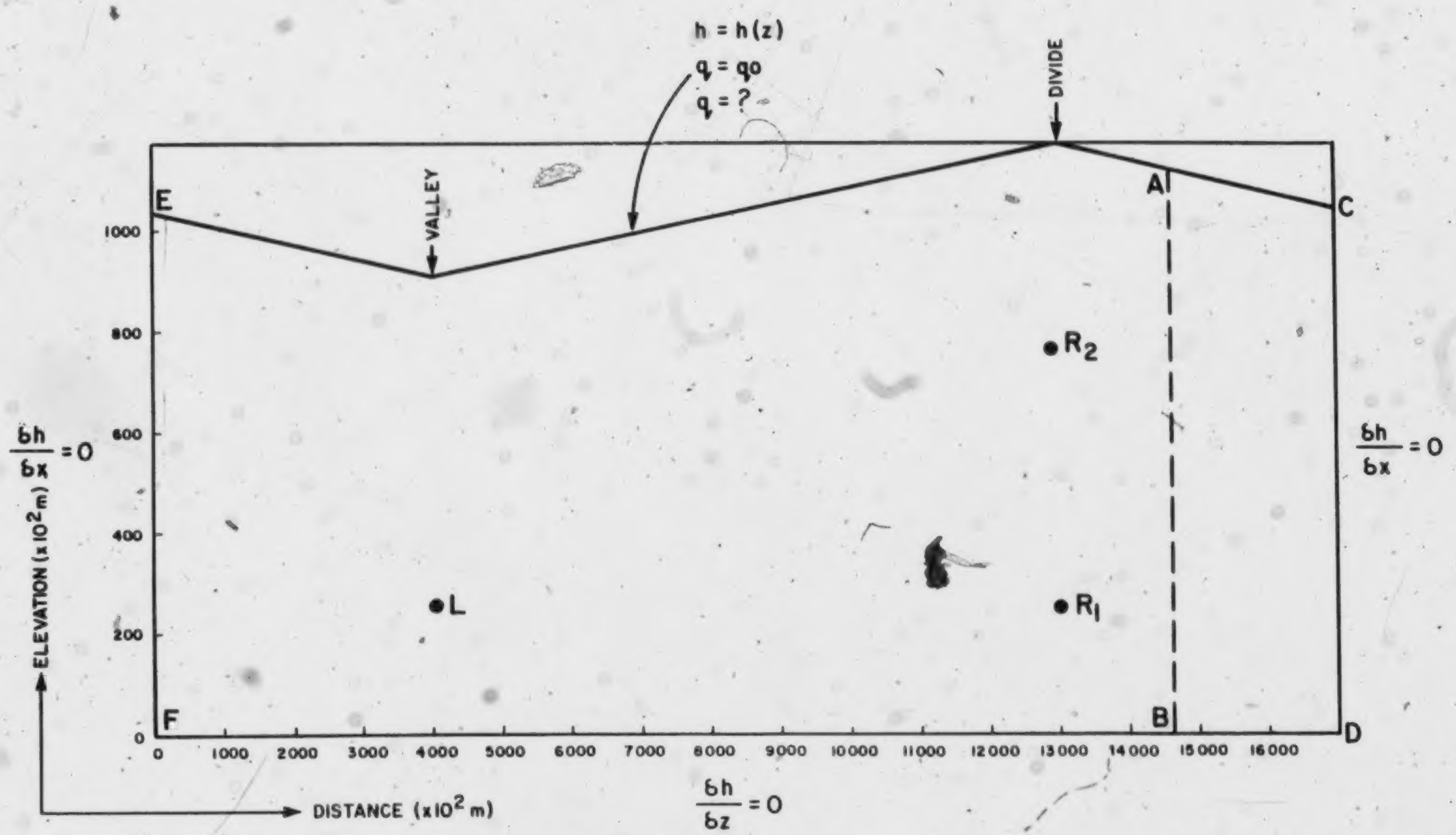

h = h(z)
q = qo
q = ?
DIVIDE
VALLEY
E
A
C
B
D
F
R2
R1
L
$\frac{\delta h}{\delta x} = 0$
$\frac{\delta h}{\delta x} = 0$
$\frac{\delta h}{\delta z} = 0$
ELEVATION (x10² m)
DISTANCE (x10² m)
0
200
400
600
800
1000
0
1000
2000
3000
4000
5000
6000
7000
8000
9000
10000
11000
12000
13000
14000
15000
16000

Figure 43. Mathematical model of BH type environment.

line AB. A deep site of development was located at point $R_1$ (below the divide) by placing a negative flux at the corresponding node during the transient run. No flux was allowed into the system along the water table. The model was then run until the drawdown at $R_1$ was equal to one-third of the available drawdown (equivalent to worse case to be considered later) at that point. The initial conditions were those of the natural steady-state flow. The "cone-of-influence" due to this development was traced and found to intercept line AB, thus, this position was unacceptable for the boundary. In subsequent runs, line AB was moved gradually toward CD and the cone-of-influence was outlined after each run. Each time the cone intercepted line AB, the boundary was rejected. At line CD, the chosen boundary, no drawdown was observed in response to development. To confirm this boundary further, several values of flux covering the estimated range of precipitation were allowed into the system at the water-table boundary in different runs. The vertical boundary was unaffected by drawdown during these runs. Finally, a shallow site of development was placed at $R_2$ and most of the earlier runs were repeated. In all cases, the boundary was not affected by drawdown, thus, an impermeable boundary was located along CD throughout the simulations. The left vertical boundary was treated in a similar manner. In this case, the site of development was located below the valley at point L. A suitable position for the boundary was found at EF (Figure 43).

One might recognize from the preceeding paragraph that the simulated site of development is a point sink in a two-dimensional cross section. The point sink corresponds to a line sink in a three-dimensional basin parallel to the basin's thalweg. The line sink is assumed to simulate a line of wells in the third dimension perpendicular to the two-dimensional cross section as will be justified later (Section 4.4).

The BH type environment was adopted for analysis in this chapter for the following reasons: 1. it is the second major (B-) category of the commonly occurring type environments in the Plains Regions, the first being the complex A-category that was analyzed in the previous chapter; 2. at a smaller scale, the B-category of basins corresponds to the smallest and incremental sub-basins of the A-category, therefore, its use here corresponds to the analysis of smaller units of the A-category; 3. it is equivalent to the unit groundwater basin as described by Tóth (1984), therefore, a unit pattern of regional groundwater flow (Tóth, 1984), consisting of the three regional groundwater flow regimes (discharge, midline, recharge) required for analysis herein could be expected to develop.

The mathematical expression of the impermeable boundary conditions in terms of hydraulic potential is the same as described in Figure 25. However, along the water table, neither hydraulic head nor flux was specified as the determination of the actual boundary condition was a part of the solution to the problem. It was only necessary to specify a maximum permissible flux along the water-table boundary.

To complete the boundary value problem for the domain and boundary conditions above, the two-dimensional equivalent of (6), the partial differential equation which describes the transient flow of water through porous media was adopted. A digital computer was then employed in obtaining a numerical solution to the formulated problem.

### 4.3 Method of solution

The finite element numerical technique for solving the equations of groundwater flow was also used in this chapter. However, the general nature of the problem at hand, and in particular, the special treatment of the water table boundary required a model that is well suited for the solution. It was required in some instances that the fluctuation of the water table be traced, therefore, a free surface model was preferred. A free surface model is capable of calculating both the unknown hydraulic head distribution and the water table configuration with time.

The model by Davis and Neuman (1983) was used for the simulations in this chapter because of its particularly adaptive features to the present problem. Essentially, only one plotting sub-routine was added to the program in terms of modification to handle the large number of simulated results. Termed UNSAT2, the model has been developed over a number of years through a series of papers. The original documentation is given in Neuman et al. (1974). The theory is described fully in the papers by Neuman (1973, 1975). The model has been applied successfully in the solution of various flow problems (Feddes et al., 1974; Kroszynski and Dagan, 1975; Wei and Shieh, 1979; Zaslavsky and Sinai, 1981).

The method of approach to the problem by this model involves that of modeling both the saturated and unsaturated parts of the groundwater system simultaneously. This method is capable of dealing with the continuum of pressure heads that exist in and between the saturated and unsaturated zones, therefore, the position of the free surface can be defined without considering it to be a boundary within the system.

A computer code which solves the pressure head equivalent of (6) for variably saturated porous media (Davis and Neuman, 1983) was used in the simulation of the transient flow problems. In particlar, the equation that is solved by this model may be written as (Neuman et al, 1974):

$$L(h)=\sum_{i=1}^{3}\sum_{j=1}^{3}\frac{\partial}{\partial x_i}\left[K^r\ (h)K^s_{ij}\ \frac{\partial h}{\partial x_j}\right]+\sum_{i=1}^{3}\frac{\partial}{\partial x_i}\ K^r(h)K^s_{j3}$$

$$-\left[C(h)+BS_s\right]\ \frac{\partial h}{\partial t}\ -\ S\ =\ 0 \qquad (22)$$

where: $L$ = quasilinear differential operator defined in flow region;

$x_i$ = spatial coordinates ($i=1,2,3$ with $x_3$ the vertical);

$K^r$ = relative hydraulic conductivity ($0 \leq K^r \leq 1$);

$K^s_{ij}$ = hydraulic conductivity tensor at saturation;

$h$ = pressure head;

$C$ = specific moisture capacity = $\frac{d\theta}{dh}$

$\theta$ = volumetric moisture content;

$\beta$ = 0 in unsaturated zone and 1 in saturated zone;

$S_s$ = specific storage;

$t$ = time; and

$S$ = sink term.

The initial conditions are described by specifying pressure heads. Commonly, boundary conditions are established by prescribing pressure head or normal flux along most boundaries of the flow region. In addition to these two boundary types, a third type of boundary can be handled by this model. This third type of boundary that has been exploited by the use of this model in the present study, results from the fact that partially saturated flow regions may be bounded by two other types of atmospheric conditions. Water may leave the system through evaporation or enter the system through infiltration at the soil-air interface. The potential rate of evaporation is dependent on atmospheric conditions, and also on the moisture content of the soil, that is, the actual rate of evaporation may be limited by the ability of the soil to transmit water upward from below. Similarly, the potential rate of infiltration may be greater than the rate at which the soil can transmit water downward and away from the boundary. For example, during a rainfall event, the precipitation rate may exceed the infiltration capacity and ponding or runoff may occur. In both cases, the potential flux at the boundary is controlled by external conditions while the actual flux depends on antecedent soil moisture conditions. Under these conditions, it is impossible to predict the exact boundary condition to be specified a priori. The UNSAT2 program provides a solution by maximizing the absolute value of the flux subject to the requirements (Hanks et al., 1969):

$$\left| K^{r} \sum_{i=1}^{3} \left[ \sum_{j=1}^{3} K^{s}_{i_j} \frac{\partial h}{\partial x_j} + K^{s}_{i3} \right] n_i \right| \leq \left| E^{*}_{s} \right| \qquad (23)$$

and

$$h_L \leq h \leq 0 \qquad (24)$$

where: $E^{*}_{s}$ = prescribed potential surface flux (m/s);

$h_L$ = minimum allowed pressure head at the soil surface (m).

This particular attribute of the model has been employed advantageously in the solution of the present problem as will be discussed later. For further details about the model the interested reader may refer to the documentation by Davis and Neuman (1983).

### 4.3.1 Standard of measurement

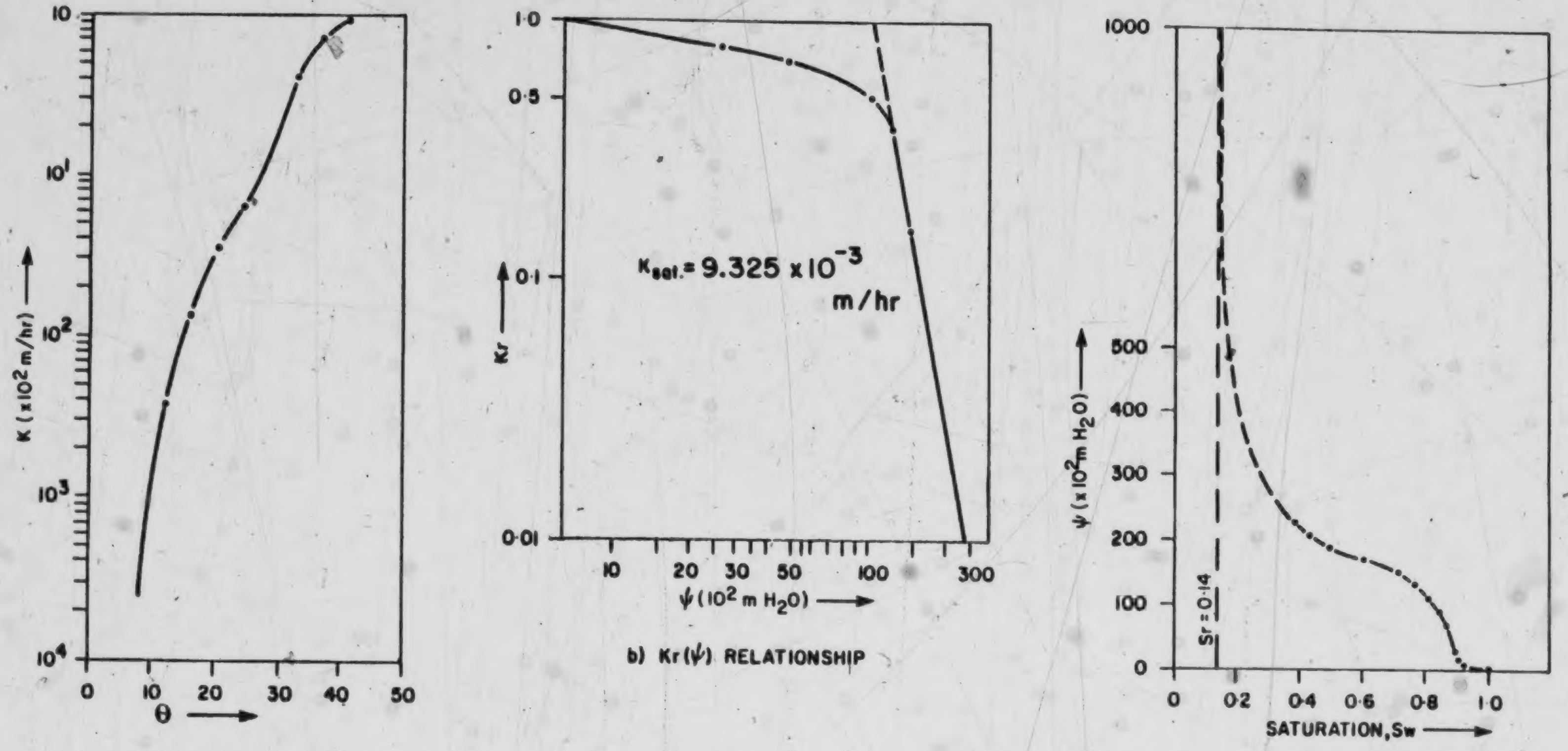
Comparison of two or more substances or situations requires that a relevant standard of measurement be set. It was for this purpose that the following problem was designed. Under virgin conditions, steady state prevails in most groundwater systems, and natural recharge is equal to the natural discharge. Discharge by wells is superimposed upon the previously stable system, and it must be balanced by an increase in the recharge of the system, or by a decrease in the old natural discharge, or by loss of storage in the system, or by a combination of these (Theis, 1940; Bredehoeft and Young, 1970). The formulation of the problem is based on this premise.

The mathematical model of the BH type environment described in Figure 43 was developed by numerical simulation so that the total inflow into the system could be calculated by the time new steady-state conditions are established from the initial steady-state conditions. The assumption in the present study is that regardless of the path by which a new equilibrium is established, the effects must be reflected by the cumulative inflow into or outflow out of the system during the transient period, because, at the new equilibrium, the rate of inflow is equal again to the rate of outflow. A quantitative estimate of the cumulative inflow due to development in the period between the initial natural and new induced steady-state conditions would characterize the transient behaviour of the groundwater system, and in particular, typify the utilizable water moving through the system during this period. Thus, this cumulative inflow has been calculated and termed as the "transitional-basin-yield (TBY)". The TBY is the standard parameter for the comparison of the response of the environment to development, and for the analysis of the stability of the basin under various scenarios of development.

The initial conditions are those of the natural steady-state flow. To determine the steady-state configuration, pressure head was specified along the top (surface) boundary to create a water table that is sub-parallel to the surface. This arrangement leaves an unsaturated zone that thickens gradually from the valley to the divide, simulating the common nature of the unsaturated zone in most Plains Regions environments. Only seepage is allowed to develop in the valley. It is presumed that the presence of a stream (constant head) at the valley will bias the comparison of the effects of the basin's response to development between the recharge and discharge areas. All other boundaries are impermeable.

Geologically, the basin is homogeneous and isotropic (BH type environment, Figure 22). A saturated hydraulic conductivity of $2.42 \times 10^{-1}$ m/day corresponding to the Paskapoo Formation was used. The unsaturated zone was more difficult to characterize because data are not collected for this zone routinely in most groundwater studies. However, some studies in the Plains Regions (Maclean, 1974; Sophocleous, 1978) have reported data for the unsaturated zone which show similar trends. These data were examined, and the set indicated in Figure 44 was selected for use in this and other simulations in this chapter. The selection of this set of data was based on the fact that the saturated hydraulic conductivity at this site is similar to that of the Paskapoo Formation. A porosity value of 0.34 and specific storage of 0.05 were used, as these values are typical of materials in the region.

The overall width of the basin is 170 m but the distance from the valley to the divide is 90 m. The depth to the horizontal impermeable boundary is 9 m below the valley. These dimensions were selected to simulate a ratio n of 0.1. Such a small basin was used so as to facilitate the modeling of the unsaturated zone. For example, the capillary fringe can be



a) K(Θ) RELATIONSHIP

b) Kr(ψ) RELATIONSHIP

c) ψ(Sw) RELATIONSHIP

Figure 44. Characteristic properties of the unsaturated zone.

Source : Sophocleous, 1978

represented on this scale, whereas, on a large scale, this zone is more difficult to represent. Furthermore, it should be reemphasized here that it is the ratio n of the environment that controls regional groundwater conditions rather that the width or depth alone.

It is possible to apply similitude considerations in extrapolating the results from the simulated regions to larger basins. Similitude considerations of Verma and Brutsaert (1970) have been developed into relationships by Freeze (1971) that may be written as:

$$\frac{x_1}{x_2} = \frac{y_1}{y_2} = \frac{z_1}{z_2} = \frac{t_1}{t_2} = \frac{h_1}{h_2} \qquad (25)$$

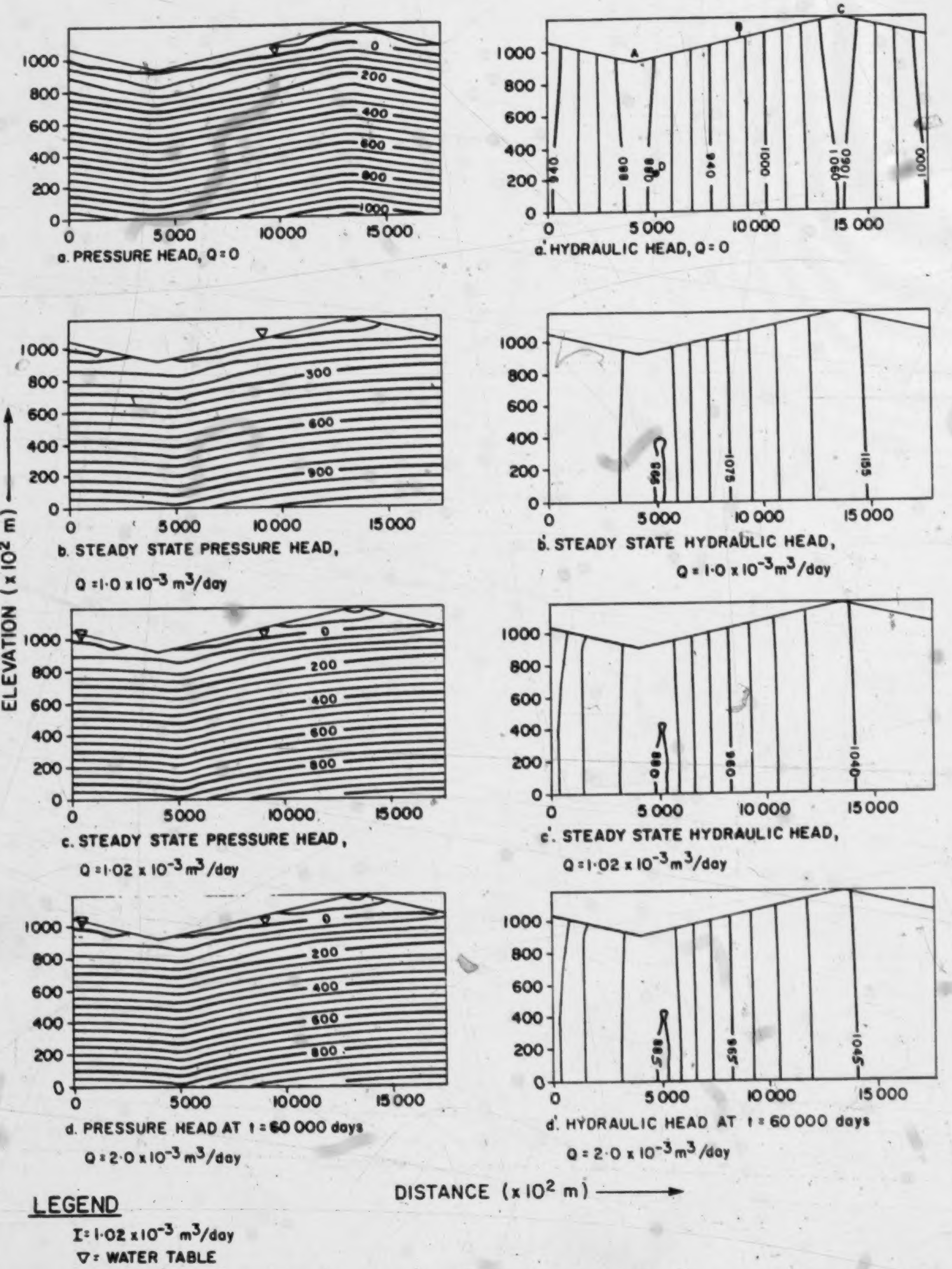
where: $x, y, z$ = length unit (m);

$t$ = time unit (s);

$h$ = head value (m).

It is obvious from (25) that extrapolation involves an interaction of length unit, time unit, and material of porous medium. Thus, similitude considerations may be extended to include pumping rates and TBY. The extent to which extrapolation can be done is controlled by a realistic range of sizes of Plains Regions basins. The range of possible extrapolation is also controlled by the scale at which soil properties and capillary fringe sizes are not altered unrealistically. The extrapolation procedure is described in the paper by Freeze (1971).

The initial steady-state flow conditions resulting from the specified boundary conditions are shown in Figures 45a, 45a', 46a and 46a'. The pressure head field is the direct numerical solution; the hydraulic head field is deduced from it. The sections have been exaggerated vertically by a factor of 8. The position of the water table is determined by the line along which pressure head is zero (Figures 45a and 46a). The hydraulic head distribution is such that a "basic pattern of regional groundwater flow" (Tóth, 1984) occurs between A and C (Figures 45a' and 46a'). Relative to the water table, groundwater flow is ascending between A and B, lateral across a vertical line below B, and descending between B and C. The distinction of these zones is based on the signs of the nodal fluxes across the surface which are calculated and displayed directly as model results. Therefore, a discharge area develops between A and B, a midline lies below B, and a recharge area characterizes the zone between B and C, under initial natural steady-state flow conditions.



a. PRESSURE HEAD, Q = 0

a'. HYDRAULIC HEAD, Q = 0

b. STEADY STATE PRESSURE HEAD, Q = 1·0 x $10^{-3}$ $m^3$/day

b'. STEADY STATE HYDRAULIC HEAD, Q = 1·0 x $10^{-3}$ $m^3$/day

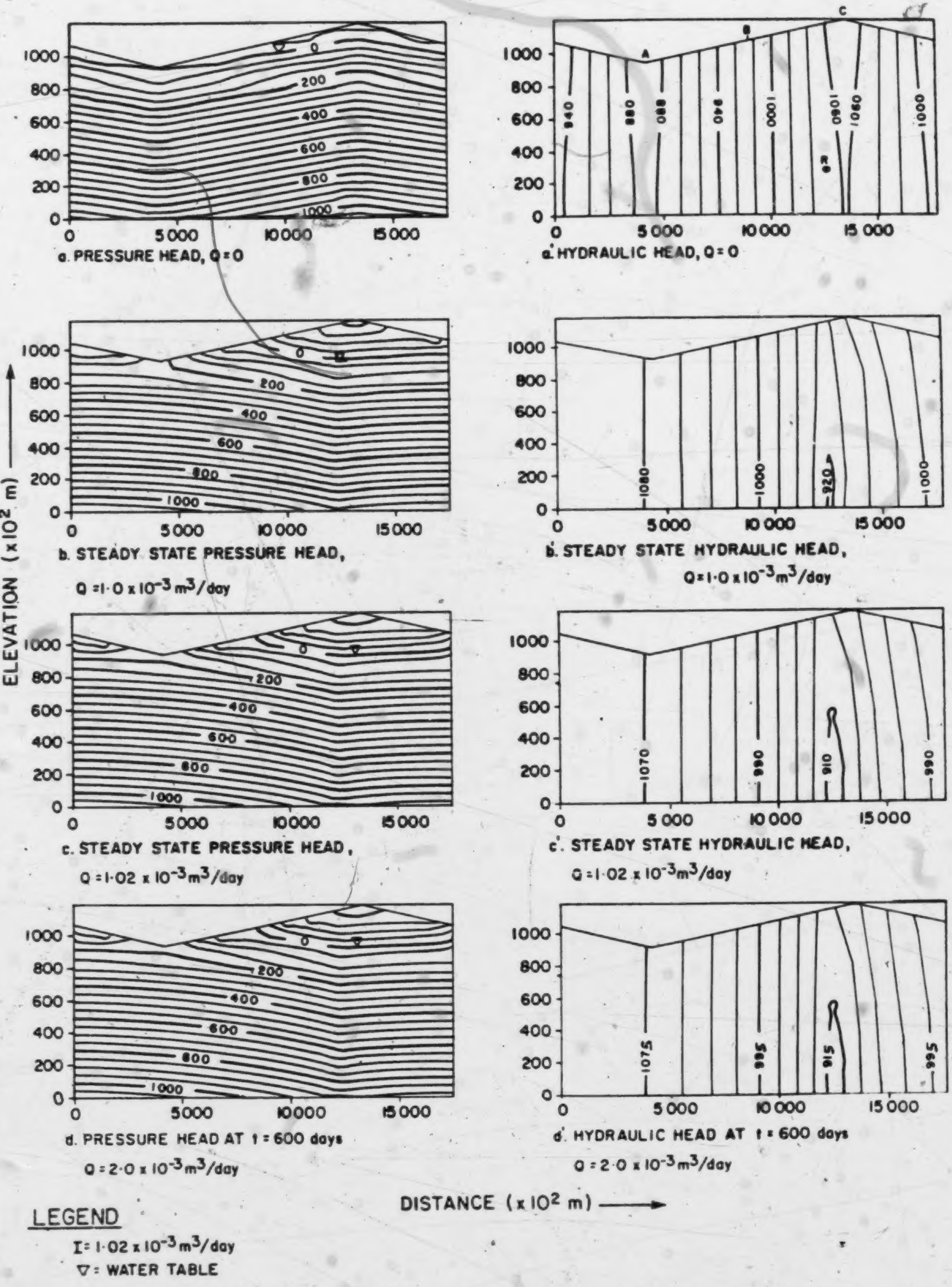
c. STEADY STATE PRESSURE HEAD, Q = 1·02 x $10^{-3}$ $m^3$/day

c'. STEADY STATE HYDRAULIC HEAD, Q = 1·02 x $10^{-3}$ $m^3$/day

d. PRESSURE HEAD AT t = 60 000 days Q = 2·0 x $10^{-3}$ $m^3$/day

d'. HYDRAULIC HEAD AT t = 60 000 days Q = 2·0 x $10^{-3}$ $m^3$/day

Figure 45. Response of BH type environment to development in discharge area with known surface infiltration rate.



a. PRESSURE HEAD, Q=0

a'. HYDRAULIC HEAD, Q=0

b. STEADY STATE PRESSURE HEAD,
Q = 1·0 x 10$^{-3}$ m$^3$/day

b'. STEADY STATE HYDRAULIC HEAD,
Q = 1·0 x 10$^{-3}$ m$^3$/day

c. STEADY STATE PRESSURE HEAD,
Q = 1·02 x 10$^{-3}$ m$^3$/day

c'. STEADY STATE HYDRAULIC HEAD,
Q = 1·02 x 10$^{-3}$ m$^3$/day

d. PRESSURE HEAD AT t = 600 days
Q = 2·0 x 10$^{-3}$ m$^3$/day

d'. HYDRAULIC HEAD AT t = 600 days
Q = 2·0 x 10$^{-3}$ m$^3$/day

LEGEND

I = 1·02 x 10$^{-3}$ m$^3$/day
∇ = WATER TABLE

Figure 46. Response of BH type environment to development in recharge area with known surface infiltration rate.

During the transient runs that followed, a constant infiltration rate of $1.02\text{x}10^{-3}$ m$^3$/day was allowed to replace the specified heads along the top boundary. In the first run, a "well" located in the discharge area (point D, Figure 45a') was pumped at a constant rate of $1.0\text{x}10^{-3}$ m$^3$/day until steady state was attained in the system. The well was simulated by allowing a negative flux equivalent to the pumping rate at the node which corresponds to point D. In the second run, a well of the same capacity was pumped in the recharge area (point R, Figure 46a'). The two runs above comprise a "study" in this section. Two other studies were executed with pumping rates of $1.02\text{x}10^{-3}$ m$^3$/day and $2.0\text{x}10^{-3}$ m$^3$/day. The resulting pressure head and hydraulic head distribution patterns of all of three studies are displayed in Figures 45 and 46, whereas, the calculated TBY values are shown in Table 10. TBY was calculated as the balance between total inflow into and outflow from the system so that positive or negative TBY would indicate a resultant gain or loss of water from the system due to development, respectively. These figures and table indicate that: 1. in study 1, steady state is established under the conditions of a fully saturated basin when the site of development is located in the discharge area, whereas, drawdown of the water table occurs above the well when it is sited in the recharge area; 2. in study 2, part of the basin now becomes unsaturated with the well in the discharge area but the drawdown in the water table is still less than that in which the well is located in the recharge area; 3. no steady state is attained in both runs in study 3, but the water table falls less rapidly above the well in the discharge area; 4. TBY is higher or less when the well is located in the discharge area or in the recharge area, respectively; 5. positive TBY values mark basin stability, whereas, unstable conditions due to development are recorded by high negative TBY values.

The results above may be generalized to include the following statements: (a) the development of a basin at a rate that is below or equal to the rate of supply of water into the system results in stable conditions, whereas, when the rate of production is above the rate of supply, unstable conditions of groundwater mining result, regardless of the location of the site of development within the basin; (b) there appears to be a quantifiable difference in the response of a regional groundwater system to development, depending upon the positioning of the site of development within the system: for example, relative to development in the recharge area, development in the discharge area tends to cause less water table drawdown, and to produce a greater TBY; (c) conditions of absolute stability in a regional system under

Table 10. TBY values due to varying pumping rates.

| Pumping rate (x $10^3$ m$^3$/day) | Condition of basin | Cumulative inflow (TBY) (x $10^6$ m$^3$) | |
|---|---|---|---|
| | | in discharge area | in recharge area |
| 1.0 | stable | 1 600 000 | 400 000 |
| 1.02 | stable | -1 | -11 |
| 2.0 | unstable | -7 260 000 | -10 149 000 |

development are marked by high positive values of TBY, those of the boundary between stable and unstable conditions are marked by low positive, low negative or zero TBY, while high negative values of TBY indicate absolute instability (Table 10), which is also manifested by a rapid temporal decline of the water table.

For the aforementioned reasons, the TBY is used in the following sections as a standard of measurement of the basin's production and stability under different conditions of development. Although the role of TBY (cumulative inflow) as employed in this study is new, the use of cumulative inflow into a groundwater system as a comparative tool is not completely foreign to groundwater studies. Harpaz (1971) evaluated the effects of clogging on specific injection rates in an artificial groundwater recharge system in Isreal by relating the rates to cumulative recharge volumes.

### 4.3.2 Response of basin to systematic development

With the establishment of a standard of measurement, a logical next step was to develop the basin systematically by pumping one well at a time with increasing distance from the valley. By these means, any systematic relation which might develop can be evaluated for basin planning and management purposes.

The mathematical model described in the preceding section was used, with all its initial steady-state flow conditions. In the transient runs however, some modifications were made in order to achieve the aims of the simulations. Pumping was done at each of nineteen well locations with distances ranging from 0 to 90 m from the valley, and the last well coincided with the axis of the topographic divide (Figure 47).

As mentioned earlier, the capture of water from direct precipitation as a consequence of regional groundwater development has not been analyzed anywhere to the author's knowledge. The flexibility with which the upper boundary of the groundwater system can be handled by the model employed in this study permits an evaluation of this subject. The upper boundary is treated as a "variable boundary" along which boundary conditions are unknown a priori. For this boundary, only a maximum potential infiltration rate (to simulate precipitation rates) is specified. The model is then allowed to calculate infiltration rates which satisfy the head distribution in the system with time, as pumping proceeds. The precipitation rates used in all the model runs of this thesis are those presumed to be typical of the Plains Regions

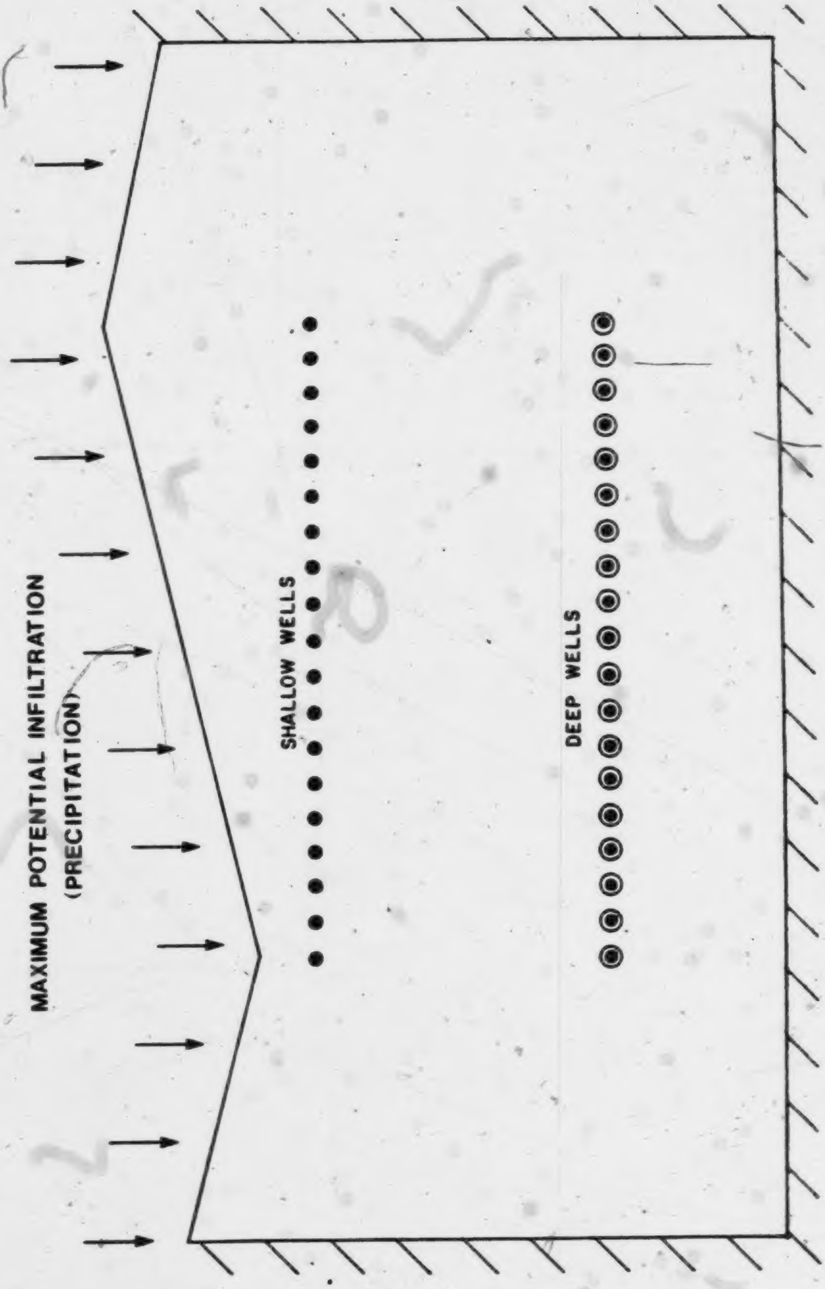

MAXIMUM POTENTIAL INFILTRATION
(PRECIPITATION)
SHALLOW WELLS
DEEP WELLS

Figure 47. Positions of pumped wells in BH type basin.

environments. The rates were estimated earlier (Section 2.3.4.1) to vary from $7.0 \times 10^{-4}$ to $1.4 \times 10^{-3}$ m/day. In some instances however, the precipitation boundaries were relaxed for convenience, and to find answers to particular questions. Throughout the analysis precipitation rate is assumed to be constant over time. This assumption may appear to be an oversimplication of the real situation, but until a site-specific study is conducted, it will be impracticable to characterize variations in precipitation rates in the Plains Regions in the necessary details.

Internal boundary conditions simulating management regulations are specified in terms of maximum allowable drawdowns at the pumping site, and in other parts of the basin. In most of the simulations, an arbitrary maximum allowable drawdown, of one-third of the available drawdown was specified. Nevertheless, the choice of any proportion of the available drawdown is a mangement decision, and its effects will be analyzed in a later section. Groundwater mining was assumed to have resulted if drawdown in any part of the basin exeeded one-third of the available drawdown either before steady state was attained, or if steady state was never attained. Throughout the analysis, groundwater mining is looked upon as an unwanted effect. The approach here and in most groundwater studies is to save groundwater in storage for generations yet unborn.

Model runs were then executed. Each run constituted pumping at each of the nineteen wells (Figure 47), until steady state was attained. To simulate "pumping" at a site, the available drawdown at the site was reduced by one-third of its value instantaneously and held constant throughout the run. With the formulation of the model in this manner, the only source of water into the groundwater system is the capture from precipitation. At steady state therefore, it is justified to assume that the calculated infiltration rate (I) across the (top) water-table boundary would be equal to the outflow rate (O). The outflow rate is equal also to the rate of production (Q) from the pumping well plus the remaining natural discharge (D) from the system. This relationship may be expressed as:

$$I = Q + D \qquad (26)$$

In this dissertation, the final steady-state infiltration rate which equals the outflow rate is referred to as the "sustainable-basin-yield (SBY)". The SBY is, therefore, the actual proportion of precipitation that is captured into the groundwater basin by development at a particular site under equilibrium conditions resulting from the development. It is the minimum rate at which the basin can be pumped at that site without mining the groundwater, every parameter remaining as defined in the model. TBY is calculated for each run simultaneously. As defined earlier, this is the total amount of utilizable water that is circulated through the basin due to development at a particular site between the initial and final steady-state conditions.

Each set of nineteen runs across the basin from the valley to the divide, in addition to the geology and boundary conditions constitute a "study". Several studies were then processed in order to analyze regional groundwater flow systematically for basin planning and management.

### 4.4 Comments on terminology and limitations

Several concepts and terms used in this thesis require clarification. For example, a "well" as used herein is a point sink in a two-dimensional, vertical cross section. In three dimensions, such a well would correspond to a line sink, representing a line of closely spaced wells parallel to the basin's thalweg and divide. A real well produces from a three-dimensional domain. A two-dimensional simulation may therefore not result in the actual three-dimensionally derived amounts of water. However, the representation of a well by a point sink in a two-dimensional cross section may be justified as follows: consider a unit basin under major groundwater exploitation with the decision to limit the siting of wells to either below (discharge area) or above (recharge area) its midline; a line of closely spaced wells could be considered to approximate these areally concentrated wells in three dimensions which may then be simulated by a point sink in two-dimensions.

Due to the large number of simulations required in this study, a three-dimensional basin-wide analysis has not been possible. Moreover, a comparative analysis involving the hydraulic regimes (discharge, midline, recharge) of a large groundwater system can be conducted easily in two dimensions, especially at this initial stage.

The calculated transitional-basin-yield (TBY) and sustainable-basin-yield (SBY) in this study apply directly to basins in two-dimensional cross sections. For an equivalent three-dimensional basin, the calculated values may be multiplied by the basin's width parallel to the thalweg and divide. However, the simulations will probably overestimate the TBY and SBY because the actual dimensionality of the problem is not accounted for completely.

Another limitation in this study is associated with the use of the terms TBY and SBY. In order to simulate natural basin hydrological conditions while maintaining appropiate boundary conditions for a unique solution to the problem, permissible drawdown has been limited arbitrarily to one-third of the available drawdown in the calcalation of TBY and SBY. In nature however, it may be necessary to develop the basin beyond one-third of the available drawdown to produce the maximum basin yield because the numerically simulated boundary conditions do not exist. In this regard, the TBY and SBY do not necessarily correspond quantitatively to the maximum basin yield.

## 4.5 Results

### 4.5.1 Presentation of results

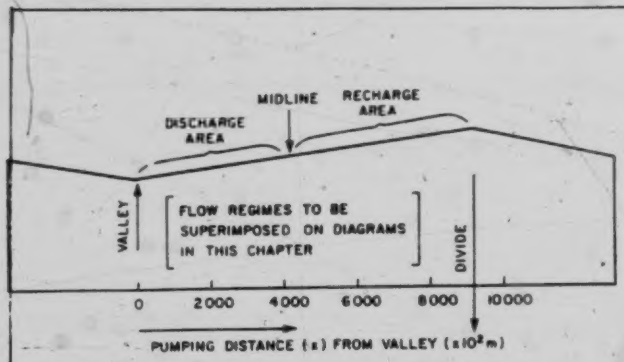
Water is needed for specific purposes in the Plains Regions (Table 7). Since the results in this chapter are presented and evaluated in the light of the uses, a brief mention of the manner of their application is necessary. Table 7 reveals that all the uses fall between two sets of end-members; namely, low volume and high rate (LH) to high volume and low rate (HL), and high volume and high rate (HH) to low volume and low rate (LL). These end-members are exemplified by oilfield activities, recreation, irrigation and domestic uses, respectively. For convenience, the interpretation of the results is based on these end-members only, and they will be referred to as use types 1, 2, 3 and 4, in that order. Furthermore, it is logical to assume that development for a use type requiring a high rate of supply should be located in areas of high SBY, that for a use of low rate in areas of low SBY, and so on. Such a plan is in agreement with the fact that optimal development of groundwater resources is not necessarily the development for maximum production, but it is synonymous with the best use (ie, the best correlation, or least discrepancy between demand type and supply type) of water.

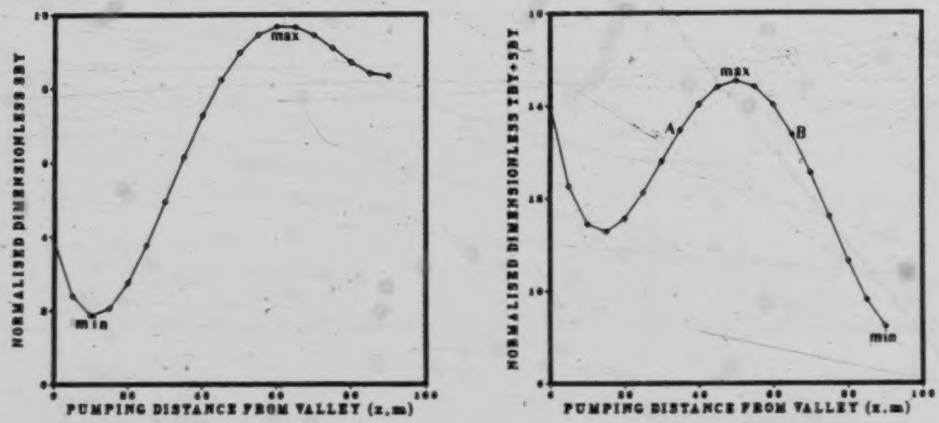
To present the results in a more explicit manner, a quantitative addition or weighting of TBY and SBY was necessary. Since these two quantities are in different dimensions, the process of addition was facilitated by an arbitrary numerical ranking procedure. In this procedure, the actual TBY values of each study are examined. The highest and lowest positive values are ranked as 10 and 1 respectively. Ranks are then calculated for the other values. The values of SBY for the same study are treated similarly. Corresponding ranks of TBY and SBY for each run of the study are then added to produce a weighted value of TBY plus SBY for each developed site in the basin. The resulting weighted values are in dimensionless units. In this form, confusion relating to units is automatically ovecome while maintaining a meaningful interpretation of the results. In a site-specific study when the actual dimensions and other hydrogeologic environmental parameters are well known, the actual calculated values of TBY and SBY may be employed directly. The use of dimensionless units here makes the approach to the study more easily applicable to the many type environments of the Plains Regions. Plots of dimensionless SBY, TBY, and SBY plus TBY versus distance of the site of development from the valley will now be presented for the several cases analyzed. The hydrogeologic environmental parameters used in the analysis are those that were determined in chapter 2 with some modifications where necessary. All the actual calculated values of SBY and TBY are shown in Appendix 2.

### 4.5.2 Main response curves

Figure 48 is a summary of the main relations that have been uncovered in this analysis. For the study described by Figure 48, the BH type environment (Figure 47) and its attributes as discussed in section 4.3.2 was used. An average regional slope of 0.03 describing the ground surface and a depth to width ratio (n) of 0.1 were used. A shallow water table of about the same regional slope was imposed by specifying the appropriate boundary conditions during the run that described the initial steady-state conditions. During the transient run, a maximum potential infiltration rate (ie. precipitation rate) of $2.42\mathrm{x}10^{-3}$ m/day was applied at the ground surface. This rate was selected to simulate both the high end of the estimated precipitation rates, as well as to coincide with a certain percentage (10%) of the saturated hydraulic conductivity. Isotropic conditions were assumed.

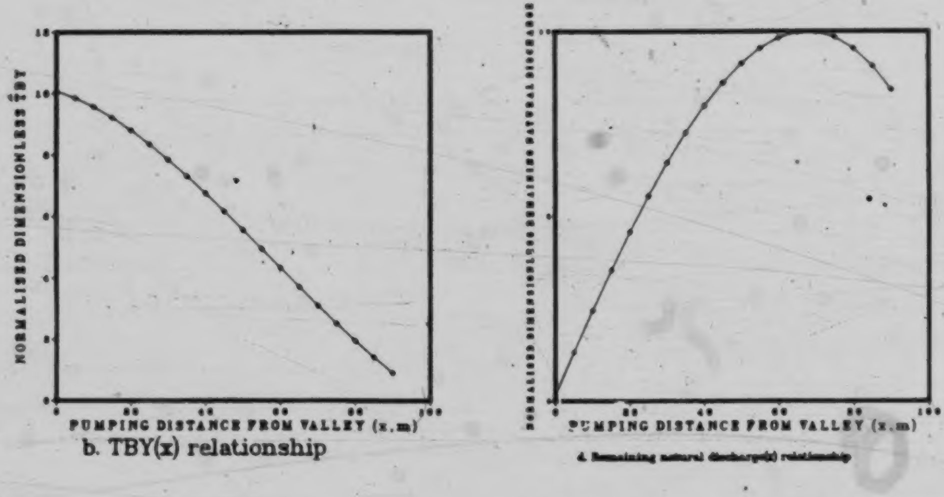
In Figure 48a, SBY is plotted against the distance of the site of development from the valley. It should be noted that the flow regimes illustrated in this figure are to be superimposed on all the related figures in this chapter. In general, SBY increases from the valley to the divide. However, points of minimum (min) and maximum (max) SBY occur at short distances away from the valley and divide, respectively, under the simulated conditions. As will be seen through the series of simulations, this seemingly apparent discrepancy is controlled by a combination of the basin's configuration and the intensity of the potential surface infiltration


MIDLINE
DISCHARGE AREA
RECHARGE AREA
VALLEY
FLOW REGIMES TO BE SUPERIMPOSED ON DIAGRAMS IN THIS CHAPTER
DIVIDE
0
2000
4000
6000
8000
10000
PUMPING DISTANCE (x) FROM VALLEY (x10² m)



a. SBY(x) relationship

c. TBY+SBY(x) relationship



b. TBY(x) relationship

d. Remaining natural discharge(x) relationship

Figure 48. Main response curves.

rate. It may be concluded from Figure 48a that, under given conditions of sufficient rainfall, the proportion of the rainfall captured into the basin at steady state by development at different distances from the valley, varies. The variation is such that minimum capture occurs with development at some point in the discharge area (near the valley), whereas, maximum capture occurs when development takes place at a point in the recharge area (near the divide). The probable explanation for this situation is that: 1. development in the discharge area preferentially uses up water moving naturally from the recharge area relative to water captured directly from precipitation; 2. in the discharge area, hydraulic gradients are naturally downward from the water table, and water moves upward toward the water table, therefore, time is required to reverse the gradient to the upward direction in order to effect downward movement and capture of water from precipitation; 3. from the recharge area, water moves away to the discharge area naturally, therefore, development in this area creates more room for water from precipitation to move into the system; 4. hydraulic gradients which are upward naturally in the recharge area are enhanced by development, causing a more rapid capture of water from precipitation. These inferences suggest that, relative to the discharge area, it is more advantageous to develop a basin in the recharge area under conditions of sufficient precipitation for use types 1 and 3 requiring high rates of water supply. Use types 2 and 4 may require that the site of development be moved toward the discharge area.

The inferences also indicate that more infiltration from precipitation is required to create equilibrium conditions by the development of a basin in the recharge area relative to the discharge area. In the absence of adequate precipitation (as is the case in some parts of the Plains Regions) therefore, the development of a basin in the recharge area may result in unstable conditions sooner than would be the case if the same basin were developed in the discharge area. This hypothesis will be investigated later.

Figure 48b is a plot of TBY versus pumping distance from the valley. According to this figure, TBY is greatest near the valley (discharge area) and decreases in an almost linear manner toward the divide. This indicates that although a little proportion of precipitation only is captured into the system by development in the discharge area, enough flow through the system from the recharge area is captured to produce a higher TBY by development in the discharge area, relative to the recharge area. Minimum TBY is obtained from development at the divide. Use types 2 and 3 are favoured by development in the discharge area relative to use types 1 and 4 which may be moved toward the recharge area.

The individual analysis of Figures 48a and 48b has shown that whereas use types 1 and 2 are favoured in the recharge and discharge areas of the basin respectively, use types 3 and 4 are equally favoured or disfavoured in both recharge and discharge areas. Since the site of development for each of these use types cannot be located in both recharge and discharge areas at the same time, an optimal position within the basin must be determined for the use types. For this purpose, a weighted curve obtained from the addition of Figures 48a and 48b as described earlier is displayed in Figure 48c, which is a tool therefore, that indicates the location in the basin at which a well may be sited with the objective of optimizing both TBY and SBY. It is clear from this figure that the maximum (max) and minimum (min) points are the best sites for the location of wells that are intended to serve use types 3 and 4, requiring high volumes and high rates, and low volumes and low rates of water supply, respectively. Where the same values occur along this curve (eg, at A and B), reference is made to Figures 48a and 48b in deciding which of the two sites is more appropriate, depending on whether a higher volume or a higher rate of supply is preferred.

Figure 48d shows the remaining natural discharge at steady state versus pumping distance from the valley. The remaining natural discharge was calculated by rearranging (26) as:

$$D = I - Q \qquad (27)$$

It is observed from this figure that pumping near the valley captures most of the water that would otherwise be discharged naturally. As pumping distance increases from the valley, more water is discharged naturally. Therefore, if the main objective of developing a basin is to capture as much of the natural discharge as possible, this figure demonstrates quantitatively, that a well in the discharge area would serve this objective better than a similar well in the recharge area. This statement is in agreement with the earlier conclusions on Figures 48a and 48b.

The observations from the response curves may be summarized as follows: 1. the positioning of the site of development within a basin is a major factor in determining the proportion of precipitation that can be captured into the basin: relative to the discharge area, a higher percentage of precipitation is captured if development is sited in the recharge area under conditions of sufficient precipitation; 2. as a result of the natural flow regimes in a groundwater basin, more water is available for use if development takes place in the discharge area relative to the recharge area; 3. a greater fraction of the natural discharge of water from a basin is captured if development is sited in the discharge area rather than in the recharge area; 4. the optimal development of a basin's groundwater resource is controlled to a large extent by the use for which the water is intended: optimal well location depends upon whether the use requires a high or low volume of supply, high or low rate of supply, or both.

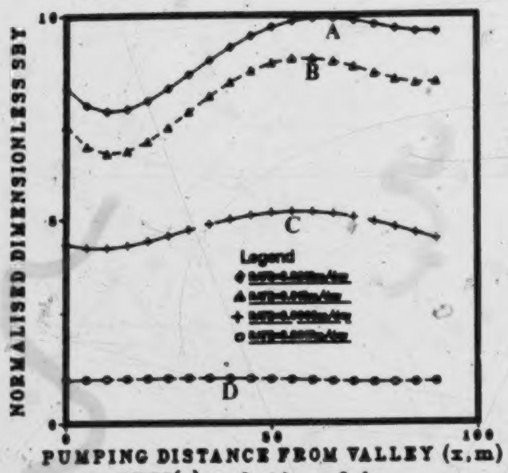
### 4.5.3 Functional relations

The major curves described in the preceding section characterize a given set of hydrogeologic environmental parameters in the Plains Regions. Since the parameters vary widely, a knowledge of the effects of their variation is also considered to be another subject of importance. A prior understanding of the role of these parameters in basin hydraulics and hydrology would improve the efficiency of exploitation. Parametrically modified forms of the curves above will now be discussed to serve the purpose of illustrating how variations in the hydrogeologic environment effect changes in the plots, and hence changes in strategies of basin planning and management. Two issues are of major concern in the presentation and explanation of these effects, namely, well location for optimal use, and development and basin stability. The discussion is subdivided on the basis of these issues.

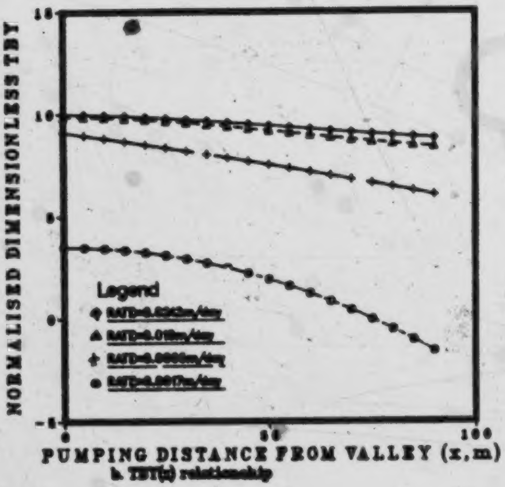
#### 4.5.3.1 Well location for optimal use

##### 4.5.3.1.1 Effect of precipitation rate

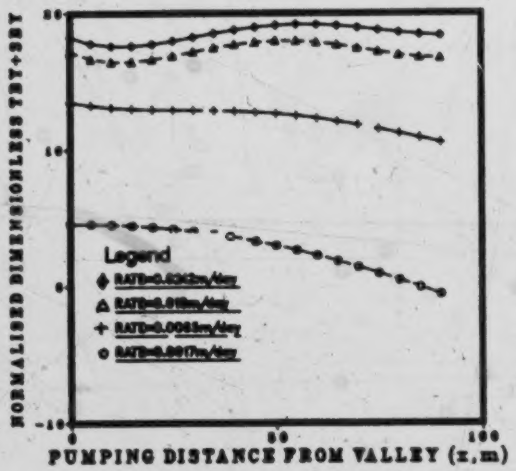
Four different maximum potential infiltration (precipitation) rates of $2.42\text{x}10^{-2}$, $1.8\text{x}10^{-2}$, $6.5\text{x}10^{-3}$ and $1.7\text{x}10^{-3}$ m/day were used in four different studies as indicated on Figure 49. The initial and boundary conditions, and all other hydrogeologic environmental parameters are the same as described for the study in section 4.4.2. The trend remains as that of increasing SBY from discharge to recharge area (Figure 49a). However, as the precipitation rate decreases, there is a tendency toward a flattening of the curves. This indicates that the distinction between the maximum and minimum SBY within a basin is greater for basins with higher precipitation rates than for those in which precipitation is less. Consequently, more planning is necessary in



SBY(x) relationship



b. TBY(x) relationship



c. TBY+SBY(x) relationship

Figure 49. Effect of precipitation on response curves.

locating a well of prescribed SBY in basins with higher precipitation than for basins with lesser precipitation. The point at which use type 1 is favoured varies with variation in precipitation rate. Whereas development for use type 1 may be sited near the divide (point A) at high precipitation rates, as the rate decreases, the point of maximum SBY (points B, C and D) moves gradually toward the valley.

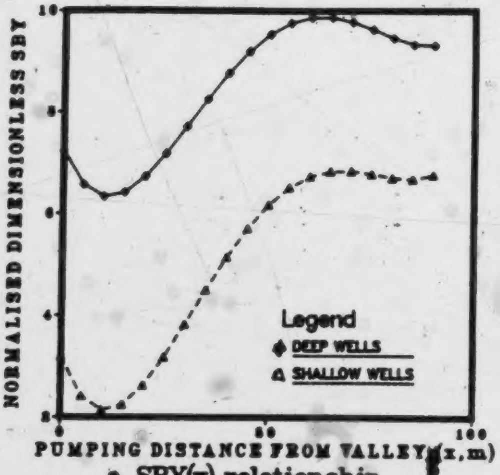
Figure 49b shows that TBY is higher if wells are located in the discharge area than with wells in the recharge area at all rates of precipitation. Contrary to the case in Figure 49a, slopes of the curves increase with decreasing precipitation rate, indicating a greater difference between maximum and minimum values of TBY as precipitation rate decreases. In all instances, use type 2 is favoured in the discharge area.

For optimal exploitation in which maximization of both volume and rate of supply (use type 3) are considered, the best well location moves from the midline area toward the discharge area for high and low precipitation rates, respectively. For minimal volume and rate of supply (use type 4), the most suitable location is at the divide for all values of precipitation considered.

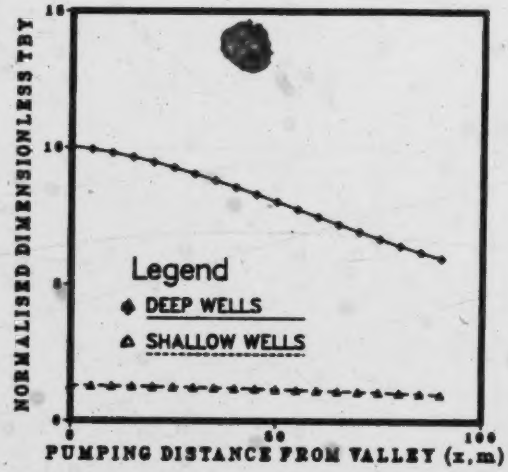
#### 4.5.3.1.2 Effect of well depth

The effect of well depth was determined by comparing the results of pumping the shallow and deep wells (Figure 47) in two studies. Figure 50 shows that changes in well depth do not lead to any displacement in the horizontal positioning of wells for optimal use. However, SBY (Figure 50a) and TBY (Figure 50b) are greater for the basin if deeper wells are developed. Also the difference between TBY in the discharge and recharge areas is more exaggerated in the deeper case. In both cases, use types 1 and 2 are favoured in the recharge and discharge areas, respectively.

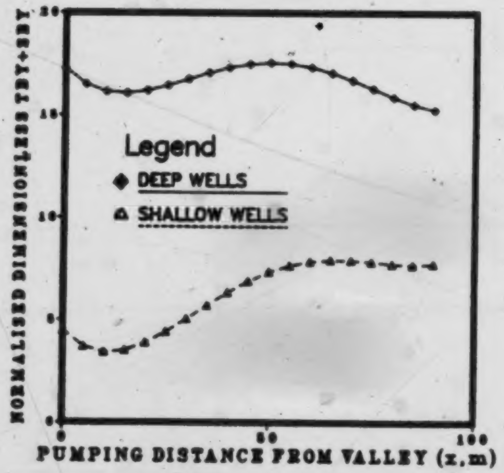
Maximization of volume and rate of supply (use type 3) requires that wells be located in the midline areas in both shallow- and deep-well cases (Figure 50c). On the criterion of minimum volume and rate (use type 4), shallow wells may be developed in discharge areas while deep wells attain their minima at the divide. For the maximization of both volume and rate, the development of deep wells is preferred.



a. SBY(x) relationship



b. TBY(x) relationship



c. TBY+SBY(x) relationship

Figure 50. Effect of well depth on response curves.

#### 4.5.3.1.3 Effect of regional slope

Three different studies were executed with the set-up in Figure 47, but with average regional slopes of 0.05, 0.03 and 0.0016 in both the initial steady state and transient runs. Only the deep wells were pumped during the transient runs.

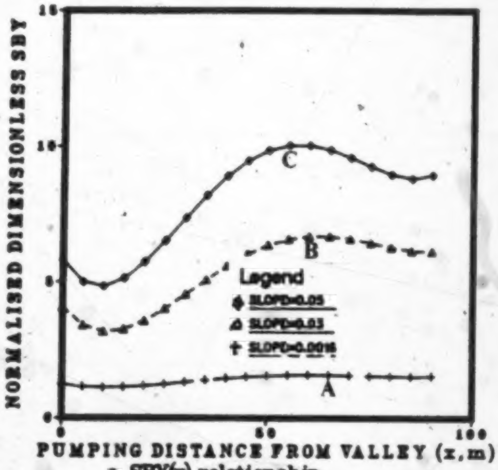
Increase in slope causes a greater difference in minimum and maximum SBY in a basin (Figure 51a). Thus, more care must be exercised in locating a well in a basin with a high slope than for a basin with a lower slope, if the main objective is to optimize SBY. Also, the point of maximum SBY is displaced further away from the divide with increasing slope (points A, B and C). Whereas use type 2 requiring high TBY is favoured in the discharge area at all slopes, TBY is greatest in the discharge area of the basin with the largest regional slope and least in the recharge area of the same basin (Figure 51b). In the basin with the smallest slope considered, TBY is reduced and becomes more uniformly distributed across the basin. This situation is expected as a high slope will naturally enhance gravity drainage of more groundwater from the recharge to discharge area of the basin, relative to a lower slope.

On the criteria of volume and rate of supply, points of maxima lie to the midline areas in general, but are displaced further away from the divide with increase in slope. That is, for use type 3, a well would have to be located at increasing distances from the divide as regional slope increases. All minimum points (for use type 4) occur in the vicinity of the divide (Figure 51c).

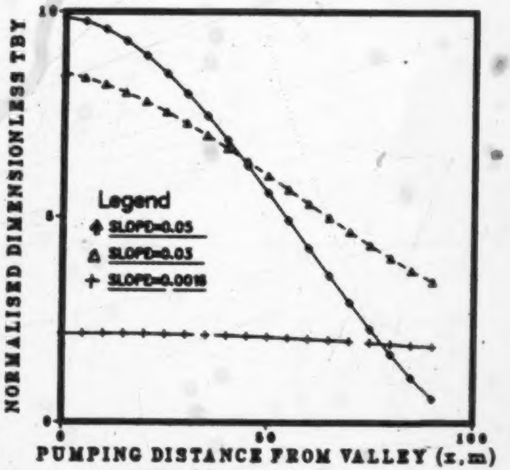
#### 4.5.3.1.4 Effect of basin depth

For this study, the depth to width ratio n of the basin was decreased from 0.1 to 0.02, every other parameter being equal. The values of SBY and TBY are higher for the deeper basin than for the shallower basin (Figures 52a and 52b). Also, there is a greater difference between the minimum and maximum values of TBY in the deeper basin. Use types 1 and 2 are favoured in the recharge and discharge areas, respectively.

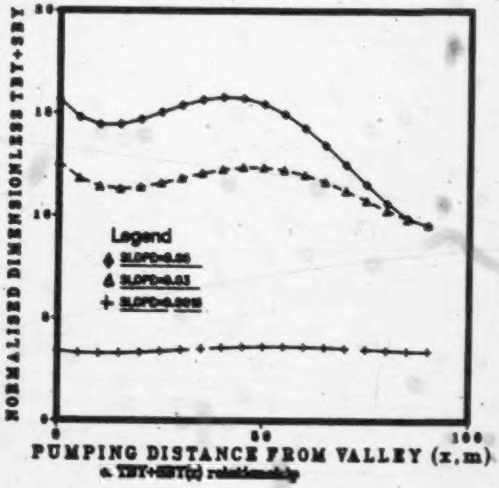
Maximization of both volume and rate indicates that wells be placed in the midline areas for use type 3 in both cases. Whereas the minimum point lies in the recharge area in the deep basin, this point lies in the discharge area in the shallow case (Figure 52c). Thus, the effects of both basin depth and well depth appear to be parallel to each other. Development in a deep basin produces more groundwater than development in a shallow basin.



a. SBY(x) relationship

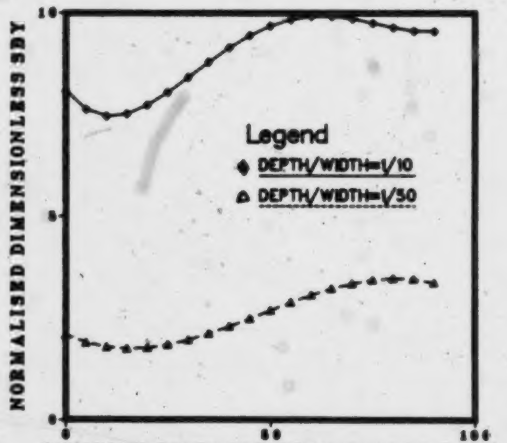


b. TBY(x) relationship

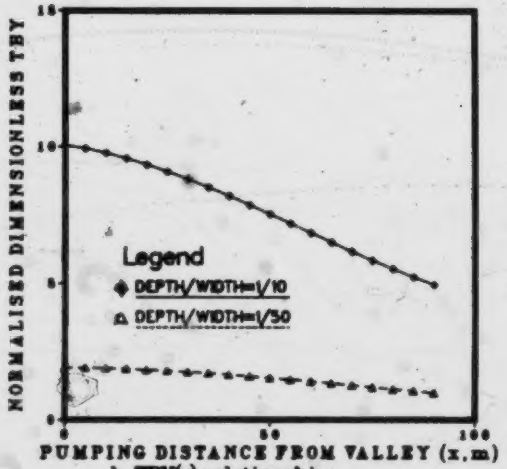


c. TBY+SBY(x) relationship

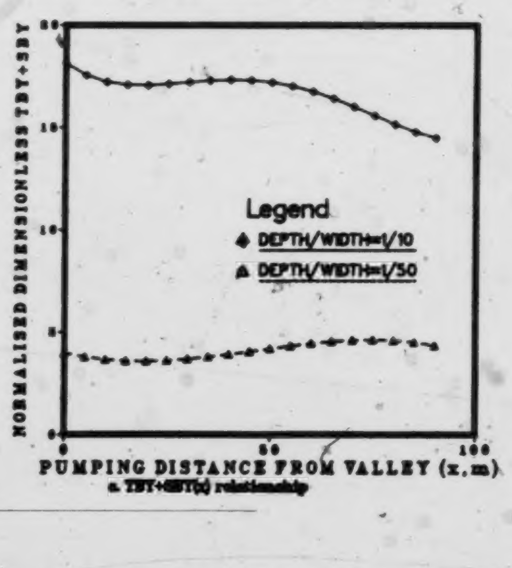
Figure 51. Effect of slope on response curves.



a. SBY(x) relationship



b. TBY(x) relationship



c. TBY+SBY(x) relationship

Figure 52. Effect of basin depth on response curves.

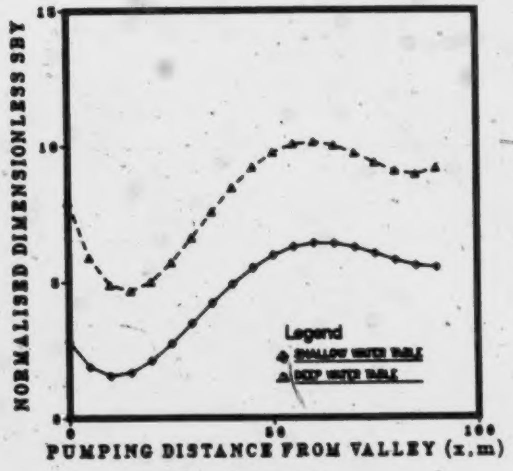
#### 4.5.3.1.5 Effect of water table depth

To determine the initial steady-state conditions for this study, pressure heads were specified along the surface (top) boundary so as to leave a water table that is deeper but parallel to that of the initial conditions of the preceding study. In all the cases, there is a correspondence of maximum and minimum points across the basins (Figure 53). The SBY and TBY values of the deep water table case are higher than those of the shallow water table basin (Figures 53a and 53b). The larger unsaturated zone above the deeper water table probably absorbs and stores more water from precipitation for onward transfer to the saturated zone under developed conditions. This situation will be possible so long as the final water table remains within the depth range in which the transfer of water from atmospheric moisture is possible. Although the difference between the minimum and maximum points is greater for the deeper than for the shallower water table case (Figure 53b), maximization of volume and rate of supply (Figure 53c) would locate wells in the midline areas of both basins. Minimization of TBY and SBY on the other hand, would locate wells in the discharge areas of both basins for use type 4. More water is produced from a basin with a deeper water table.

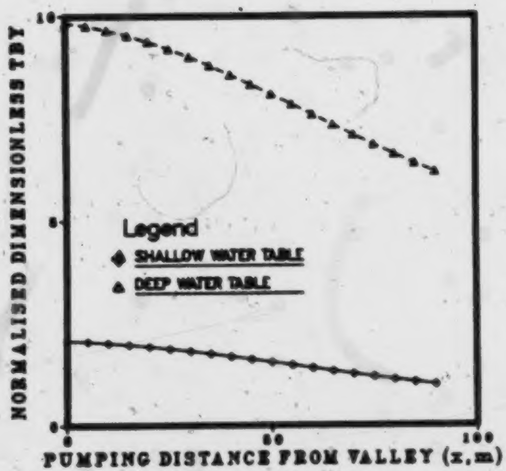
#### 4.5.3.1.6 Effect of hydraulic conductivity

Since the hydraulic conductivity of a porous medium has an important control on the ability of the medium to transmit any liquid through it, the effects of its variation in basin development deserve a consideration. Two approaches were employed: first, the whole basin was assumed to be homogeneous and isotropic, but the value of the hydraulic conductivity was varied; second, the basin was heterogeneous and either an extensive aquifer occupied a deep or shallow depth, or a restricted aquifer occurred at the discharge, midline or recharge area at a shallow depth.

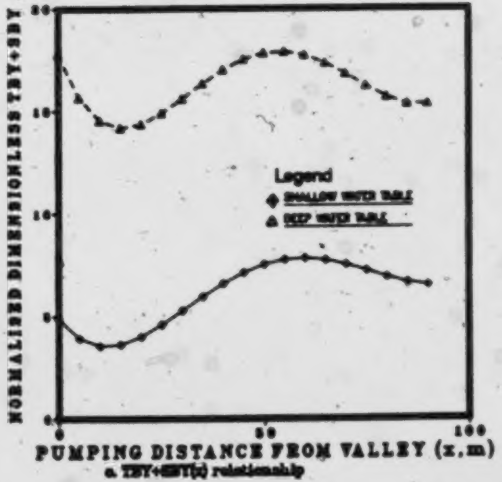
Figure 54 shows the effect of varying the entire basin's hydraulic conductivity. For the three different values of hydraulic conductivity, a regional slope of 0.03 was used, whereas, a ratio n of 0.1 and a precipitation rate of $1.8 \times 10^{-3}$ m/day were employed. In Figure 54a, the SBY values resulting from the three studies are displayed. It is obvious that the higher the hydraulic conductivity is, the higher the SBY. At a hydraulic conductivity of 24.2 m/day, the SBY attains the precipitation rate across the basin, thus, SBY is constant at every site of development (curve d). In the other two studies, points of minimum and maximum SBY lie in the discharge



a. SBY(x) relationship

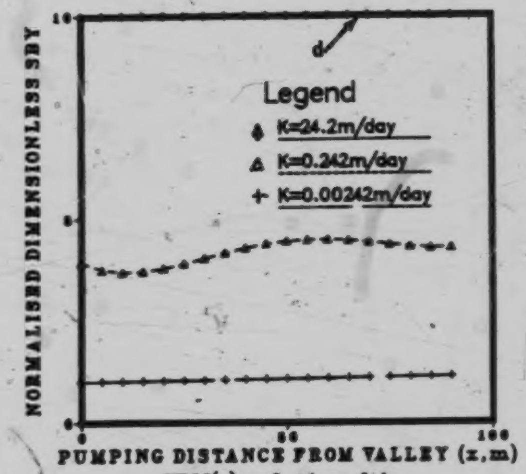


b. TBY(x) relationship

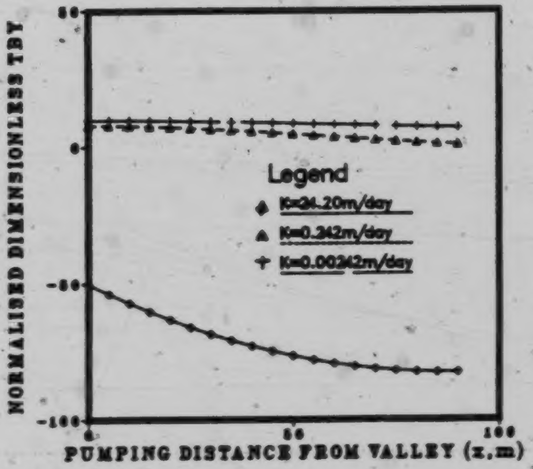


a. TBY+SBY(x) relationship

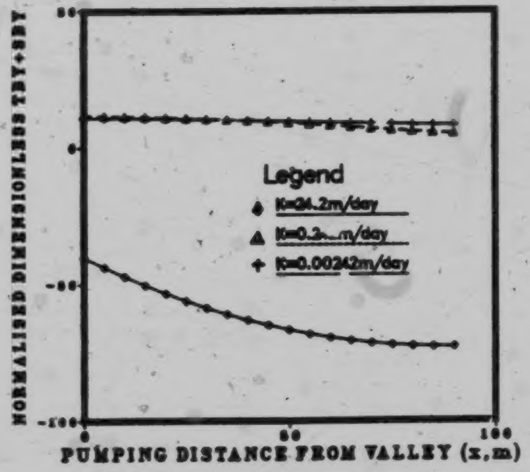
Figure 53. Effect of water table depth on response curves.



a. SBY(x) relationship



b. TBY(x) relationship



c. TBY+SBY(x) relationship

Figure 54. Effect of hydraulic conductivity on response curves.

and recharge areas, respectively. It is noteworthy that with the hydraulic conductivity of $2.42\times10^{-3}$ m/day, the minimum and maximum points of SBY coincide with the valley and divide, respectively, suggesting that the rate of precipitation relative to the hydraulic conductivity of the basin controls the location of the points of minima and maxima to a certain degree. In general, use type 1 should be sited in the recharge area.

The equivalent TBY curves of the studies above are shown in Figure 54b. This figure shows an inverse relationship between TBY and hydraulic conductivity. The inference that can be made is that: although a basin of highly conductive geologic materials will capture water from precipitation at a faster rate than that with a low hydraulic conductivity, given sufficient time, sufficient precipitation, and equal available drawdown, the basin of lower hydraulic conductivity will eventually yield a greater quantity of water at steady state, following development. Nonetheless, the rate at which the low conductivity material delivers groundwater is often much lower than is required for use. This is the main factor that precludes their exploitation as aquifers. There is a tendency toward a flattening of the curves as hydraulic conductivity decreases, indicating similarity in values of TBY and SBY that can be developed across the basin (Figure 54c). The siting of development for use type 2 will be best in the discharge areas of the basins.

The positioning of the extensive and restricted aquifers that were analyzed in the next five studies is shown in Figures 55 and 56, which represent the BTE and BTR type environments of the Plains Regions, respectively. In each case, the hydraulic conductivity of the aquifer is two orders of magnitude higher than that of the surrounding aquitard.

Figure 57a shows that a shallow extensive aquifer in a basin produces higher and lower values of SBY in the recharge and discharge areas relative to a deeper aquifer. Generally, development for use type 1 should be sited in recharge areas in both cases. Figure 57b indicates that TBY is higher in the discharge area in both deep and shallow aquifers, but the shallow aquifer is preferred for development because of its higher TBY across the basin. Maximization of volume and rate of supply also indicates an advantage in the development of the shallow aquifer relative to the deep one (Figure 57c). In both cases, use types 3 and 4 are to be located preferentially in the discharge and recharge area, respectively.

Relative to the background (ie. when basin is made up of aquitard only), the effect of aquifers is to increase SBY (Figure 58a), and to decrease TBY (Figure 58b). The more extensive

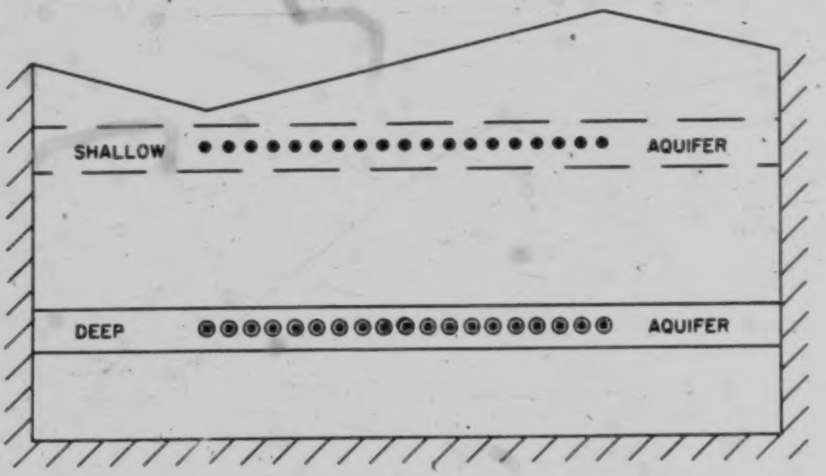

SHALLOW
AQUIFER
DEEP
AQUIFER

Figure 55. Positions of extensive aquifers in BTE type environment.

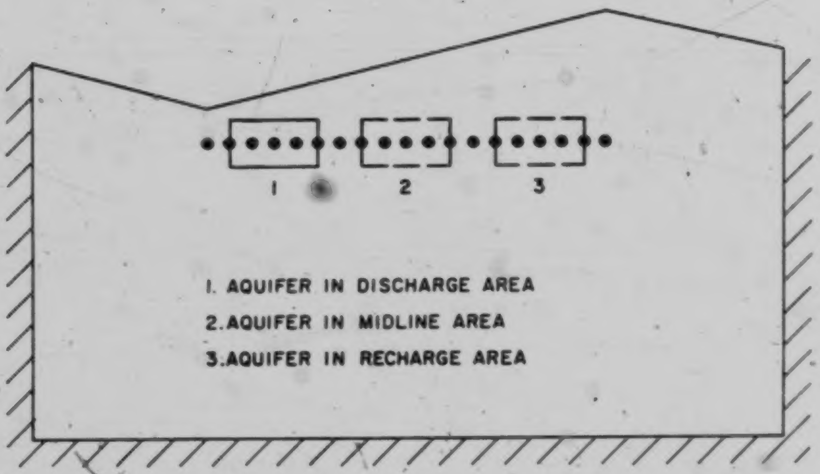
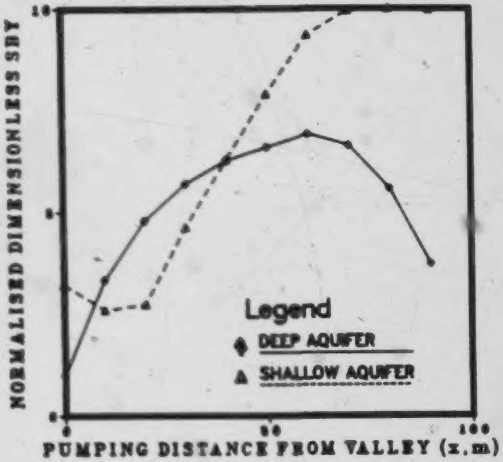
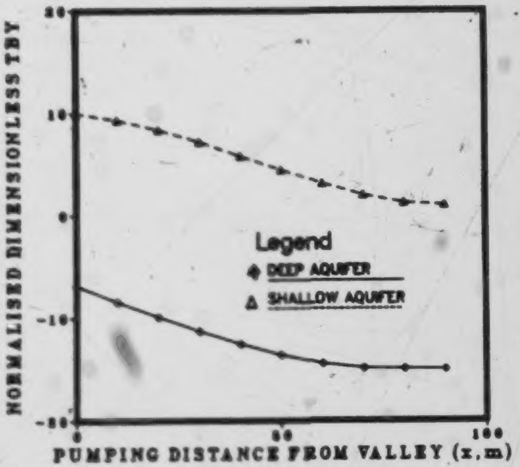

1
2
3
1. AQUIFER IN DISCHARGE AREA
2. AQUIFER IN MIDLINE AREA
3. AQUIFER IN RECHARGE AREA

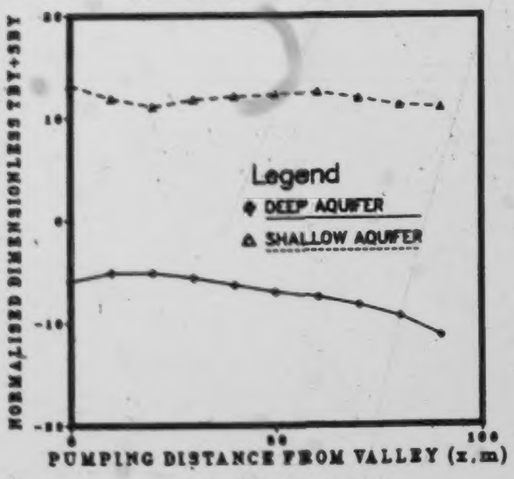
Figure 56. Positions of restricted aquifers in BTR type environment.



a. SBY(x) relationship

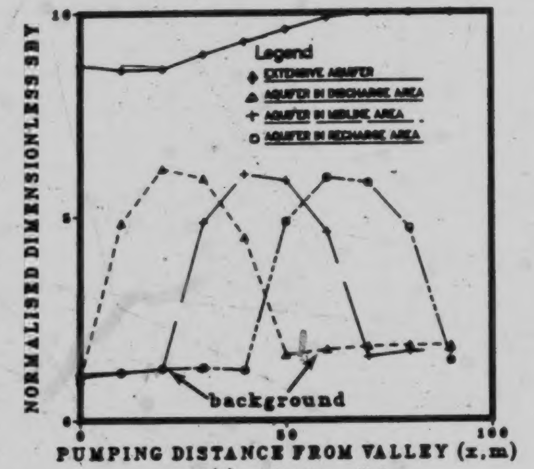


b. TBY(x) relationship

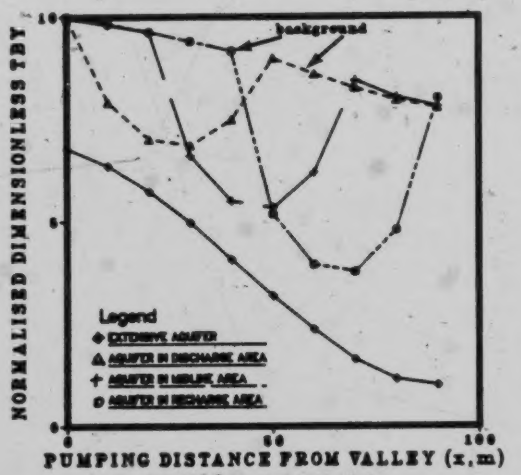


c. TBY+SBY(x) relationship

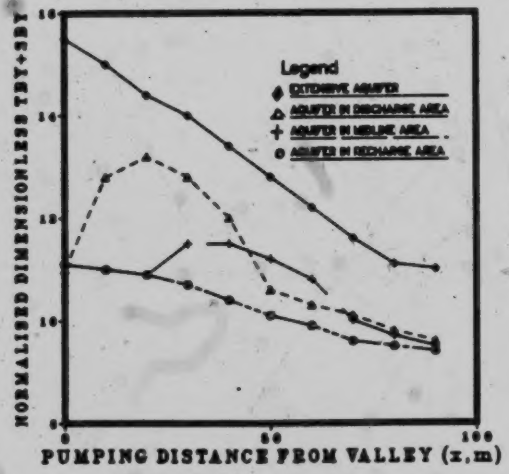
Figure 57. Effect of positioning of extensive aquifers on response curves.



a. SBY(x) relationship



b. TBY(x) relationship



c. TBY+SBY(x) relationship

Figure 58. Effect of restricted aquifers on response curves.

the aquifer is, the greater is the increase or decrease in SBY or TBY. In the case of the restricted aquifers, the effect is greater at their midpoints. Thus, the midpoint of a high permeability lens is recommended for development when the objective is to locate areas of high SBY, but this point is the least productive in terms of TBY. Also, the amount by which the SBY or TBY is increased or decreased varies with the position of the aquifer in the basin. An aquifer in the discharge area increases SBY by the greatest amount but decreases TBY by the least amount, the converse being true for an aquifer in the recharge area. The overall effect is that maximization of both SBY and TBY produces peaks at the midpoints of the aquifers, but these peaks are masked out gradually toward the recharge areas (Figure 58c). Therefore, high permeability lenses are more effective in their contribution toward the water resources of a basin when they occur in the discharge areas relative to the recharge areas.

#### 4.5.3.1.7 Summary

The uses for which water is needed in the Plains Regions of Alberta may be characterized by two sets of end-members, namely, those requiring high rate and low volume of supply (use type 1) to those requiring low rate and high volume of supply (use type 2), and those of high rate and high volume (use type 3) to those of low rate and low volume (use type 4). Their differing requirements indicate that sites of development intended to serve each of these use types in a basin cannot be located at the same spot with optimal results.

Categorization of the uses in the above manner causes a coincidence of the use types with the natural regimes of regional groundwater flow, a useful guide in locating their optimal sites of development.

Generally speaking, for optimal development, use types 1, 2, and 3 should be located in the recharge, discharge and midline areas of a basin, respectively. Use type 4 is most often favoured at the divide. Exceptions to these general rules occur with the presence of high permeability lenses in the basin, requiring that use types 1 and 3 be sited at the midpoints of the lenses, and types 2 and 4 remain at the valley and divide respectively, for optimal results.

All other controlling parameters being equal, higher SBY and/or TBY may be obtained if wells are sited: 1. in basins with higher rainfall; 2. at deeper levels in the basin; 3. in basins with higher regional slopes for SBY, and in the discharge areas of basins with higher regional slopes for TBY; 4. in deeper basins; 5. in basins with deeper water tables; 6. in materials with higher hydraulic conductivities for SBY, and with lower hydraulic conductivities for TBY; 7. in extensive aquifers rather than restricted aquifers for SBY, and in restricted aquifers rather than extensive aquifers for TBY; 8. in aquifers in discharge areas rather than in aquifers in recharge areas.

#### 4.5.3.2 Development and basin stability

Basin stability is defined here as the establishment of new steady-state conditions of flow due to groundwater development, other than the initial natural equilibrium conditions of regional groundwater flow. Although the consideration of basin stability as a condition of groundwater development has undergone a critical review by various hydrologists (Kazmann, 1956; Bear and Levin, 1967; Domenico et al., 1968), the concept may still be valid as is evident from the papers of several others (Bredehoeft and Young, 1970; Freeze, 1971; Kovács, 1981; Bredehoeft et al., 1982). A detailed discussion of the advantages and disadvantages of the concept is beyond the scope of this thesis, but the formulation of the present regional groundwater flow problem permits an evaluation of the stability of a basin under conditions of development.

As stated in section 4.3.1, stability in a basin is marked by high positive values of TBY; the boundary between stable and unstable conditions by relatively low positive, low negative or zero TBY; and instability by high negative TBY accompanied by a rapid decline of the water table. Basin stability is discussed by employing these criteria. For convenience, three stability zones were established as follows: 1. stable zone: the zone in which TBY is above zero; 2. transition zone: the zone in which TBY ranges from zero to a negative value at which a rapid decline of the water table starts to occur; 3. unstable zone: the zone of high negative values of TBY and rapid decline of water table. It was observed during the analysis that the boundary between transition and unstable zones is difficult to determine because it is not marked by a characteristic value of TBY. Consequently, emphases have been placed on the stable-transition zone boundary which is typically marked by a distinct zero value of TBY in the discussion that follows.

All conditions of simulations are similar to their counterparts in section 4.5.3.1, therefore, only the direct results are presented in this section.

#### 4.5.3.2.1 Effect of precipitation rates

Figure 59 portrays the relation between TBY and values of maximum potential infiltration (precipitation) rates for a well located in the discharge area of the modified BH type environment. The figure shows the observed trend of events at that well location during several runs with precipitation rate as the variable. Stable, transition and unstable zones were observed in the range of precipitation rates analyzed. In the stable zone, precipitation rate is above $3.8\times10^{-3}$ m/day. Steady-state conditions were established with positive values of TBY. As the precipitation rate was decreased below $3.8\times10^{-3}$ m/day, steady-state conditions were established at the expense of negative TBY. TBY became more negative gradually with further decrease in precipitation rate. Eventually, groundwater mining set in that was manifested by a rapid lowering of the water table, indicating the unstable zone. At this particular site of development, a precipitation rate of $2.0\times10^{-5}$ m/day resulted in groundwater mining, whereas, a precipitation value of $2.0\times10^{-4}$ developed equilibrium conditions. An attempt to define the actual transition-unstable zone boundary resulted in wasteful expenses and time loss. Thus, an approximate boundary only is indicated by the discontinuous line (Figure 59). Suffice it to say that decreasing precipitation rates will cause development at a particular site in the basin to move from stable through transitional to unstable conditions, every other parameter being equal.

Figure 60 shows the zones described in the preceding paragraph for two different sites of development (D in discharge area, Figure 45a'; and R in recharge area, Figure 46a'). The conclusion that can be reached from this figure is that development in the recharge area results in conditions of transition zone sooner than if development were to take place in the discharge area, as precipitation rate decreases. This difference has been marked as lag zone 1 (Figure 60). Another lag zone (lag zone 2) occurs between the transition and unstable zones from the development at the two different sites.

In Figure 61, the simulated steady-state drawdown cones due to development in the discharge area (Figure 61a), and in the recharge area (Figure 61b) are presented. These figures show that transition and unstable conditions occur at higher precipitation rates in the recharge area than in the discharge area.

Figure 62 is a further illustration of the behaviour of the basin due to development under various precipitation scenarios. Within the range of $1.8\times10^{-2}$ to $2.42\times10^{-2}$ m/day of

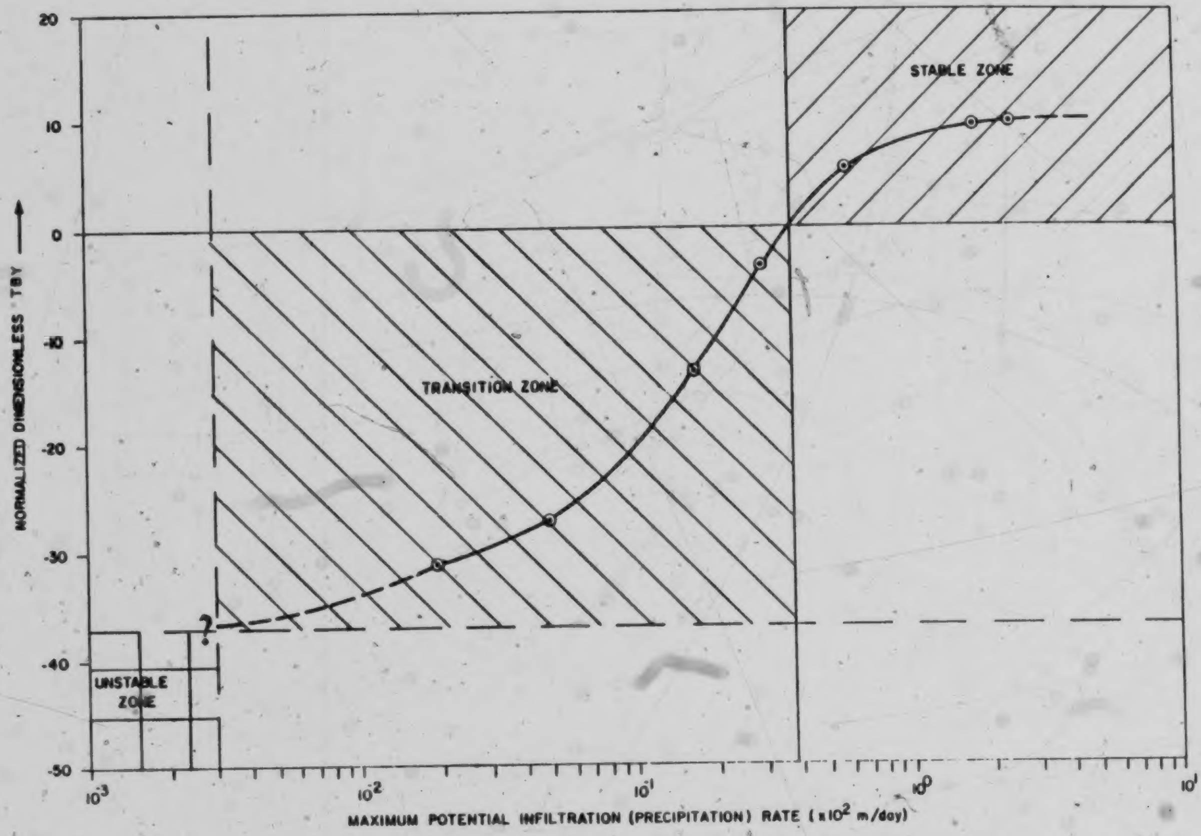


Figure 59. TBY versus precipitation rates for several runs for a pumping well in discharge area

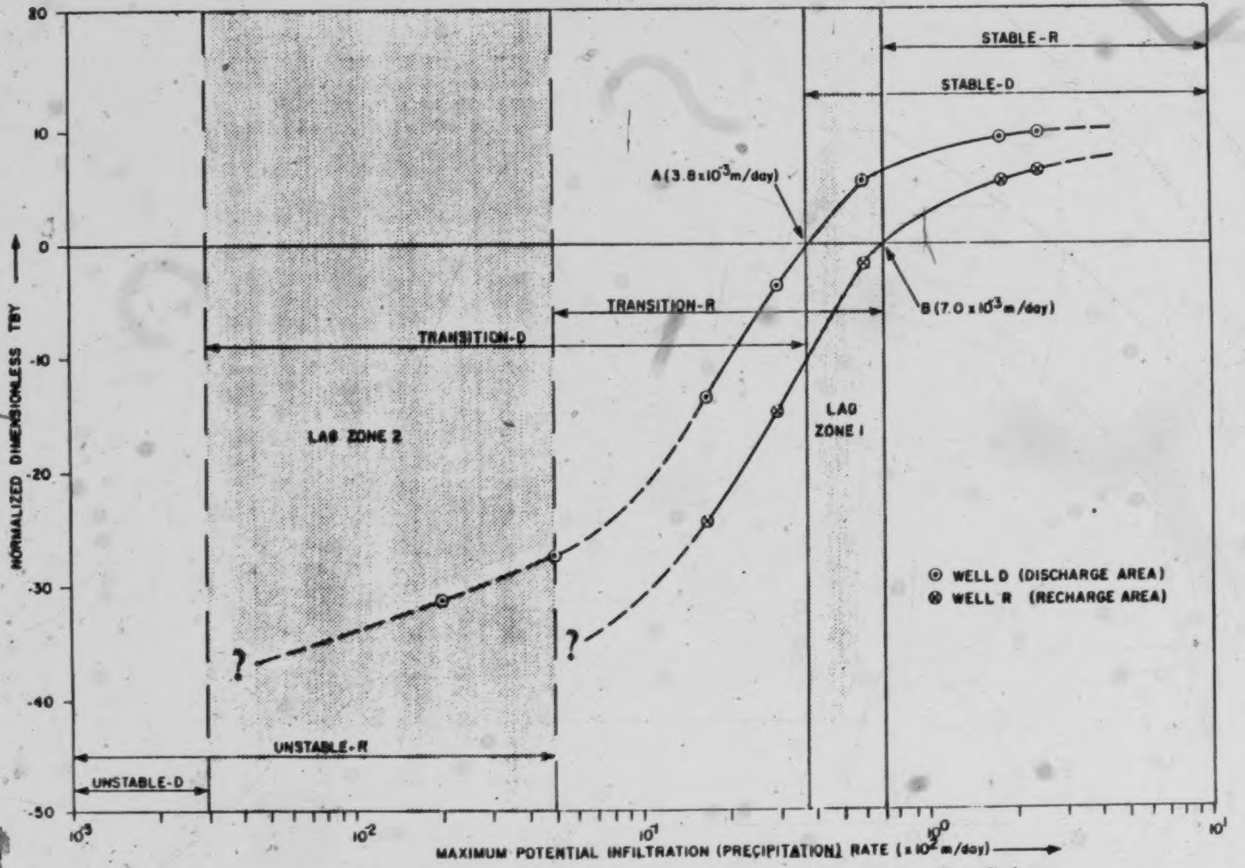


Figure 60. TBY versus precipitation rates for several runs for pumping wells in discharge and recharge areas.

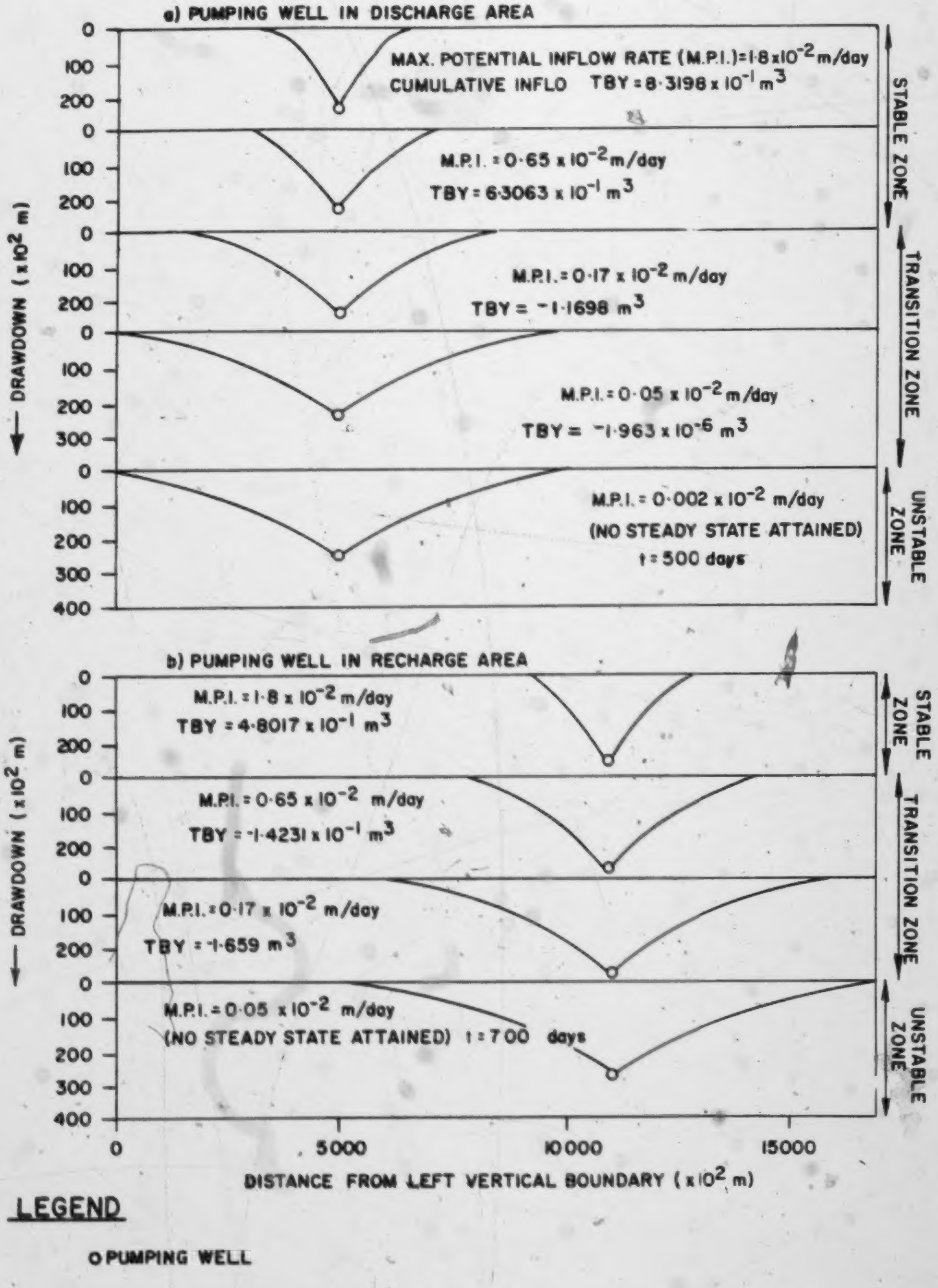


Figure 61 Steady state drawdown cones due to development under different precipitation rates.

precipitation rate, development in any part of the basin (discharge or recharge area) results in equilibrium conditions. At a precipitation rate of $6.5\times10^{-3}$ m/day, equilibrium conditions are also established in all parts of the basin, but while conditions of development in the discharge area lie in the stable zone, conditions of development in the recharge area are already in the transition zone. This situation indicates a marked difference in the ultimate response of the basin to development in either of its discharge or recharge area. With a precipitation rate of $1.7\times10^{-3}$, the basin's response to development in all parts lies in the transition zone.

Based on the inferences above it is concluded that a basin might be developed in its discharge area to produce groundwater on sustained basis, whereas, the groundwater in the same basin may be mined if the development is sited in the recharge area. This is considered as a crucial point in regional groundwater development and management.

#### 4.5.3.2.2 Effect of permissible drawdown

In two different studies, two management scenarios were simulated. In the first, the decision is to use up one-third of the available drawdown as opposed to consuming one-tenth of the available drawdown in the second. Figure 63 shows that although less water is available for use with the decision to utilized one-tenth of the available drawdown, stability is assured with development in any part of the basin. On the other hand, there is a tendency toward groundwater mining if development is sited in parts of the recharge area with the decision to use up one-third of the available drawdown in the basin.

#### 4.5.3.2.3 Effect of well depth

Whereas deep wells will enhance the availability of water for use in the discharge area relative to shallow wells, transition zone conditions will be obtained by developing the basin in the recharge area with deep wells (Figure 64). Shallow wells produce less water, but the basin maintains stability across its width.

#### 4.5.3.2.4 Effect of water table depth

Probably due to regulation from the supply in the thicker unsaturated zone, the development of a basin with a deeper water table results in stable conditions from discharge to recharge area (Figure 65). The development of the same basin with a shallower water table in

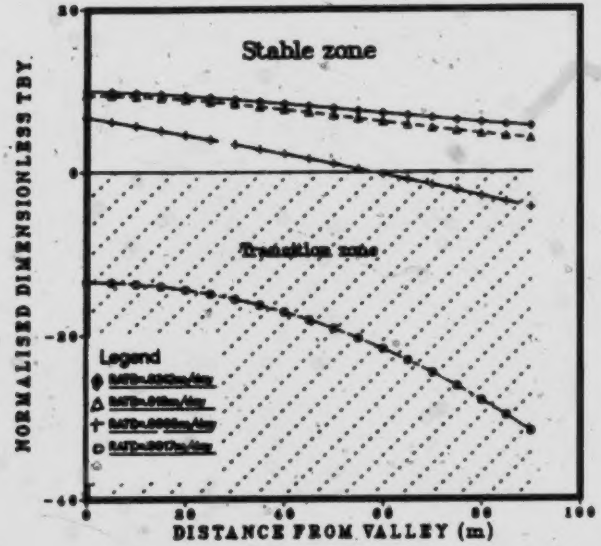


Figure 62. Effect of precipitation on basin stability.

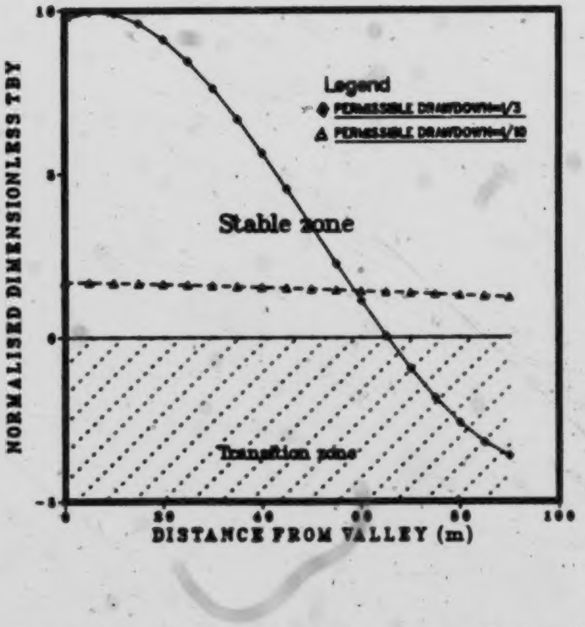


Figure 63. Effect of permissible drawdown on basin stability.

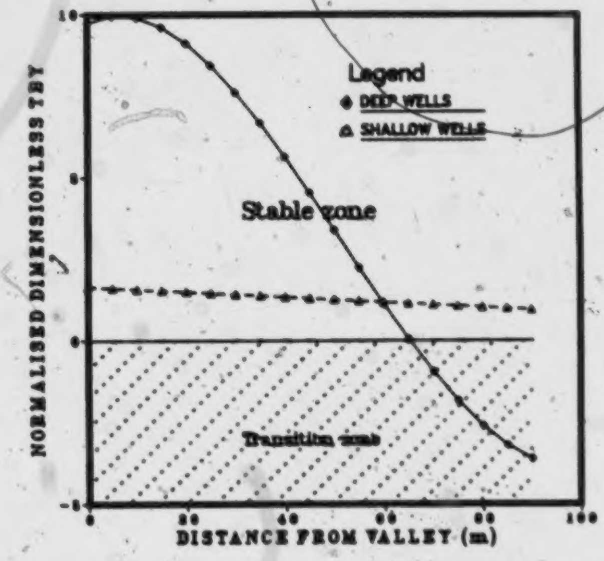


Figure 64. Effect of well depth on basin stability.

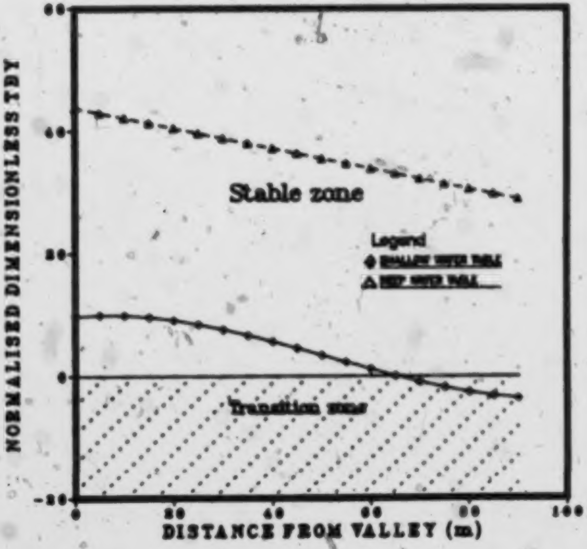


Figure 65. Effect of water table depth on basin stability.

the recharge area may tend toward instability.

### 4.5.3.2.5 Effect of basin depth

Figure 66 shows that a shallow basin would produce less utilizable water than a deep basin for a specified percentage of available drawdown. However, stability is assured during production at any site in the shallow basin, but production in the recharge area of the deep basin results in conditions that approach instability.

### 4.5.3.2.6 Effect of regional slope

Basins with low regional slopes tend to produce stable conditions during development (Figure 67). As regional slope increases, conditions of transition begin to set in from development in the recharge area. The higher the regional slope, the larger is the proportion of the basin that will produce unstable conditions during development.

### 4.5.3.2.7 Effect of hydraulic conductivity

Figure 68 indicates that the development of basins of high hydraulic conductivity materials will cause unstable conditions sooner than for materials of lower hydraulic conductivity. As will be seen in the next section however, it is the ratio of the available precipitation to that of the hydraulic conductivity that controls basin stability during development rather than either of these parameters alone.

### 4.5.3.2.8 Effect of precipitation to hydraulic conductivity ratio

As a further test of the conclusion regarding the role of hydraulic conductivity, three studies were conducted with varying values of precipitation rate and a constant hydraulic conductivity of 24.2 m/day (Figure 69). Again, conditions of development across the basin fall in the transition zone for the lower precipitation rates of $1.8 \times 10^{-3}$ and $1.8 \times 10^{-2}$ m/day. As the precipitation rate is increased to 1.8 m/day, stable conditions are restored during development across the basin. Recognizing that the ratio 1.8 to 24.2 in Figure 69 is the same as that of $1.8 \times 10^{-2}$ to $2.42 \times 10^{-1}$ in Figure 68 which produced stable conditions, it is clear that the ratio of precipitation rate to that of the hydraulic conductivity is the sole parameter that controls basin stability during development.

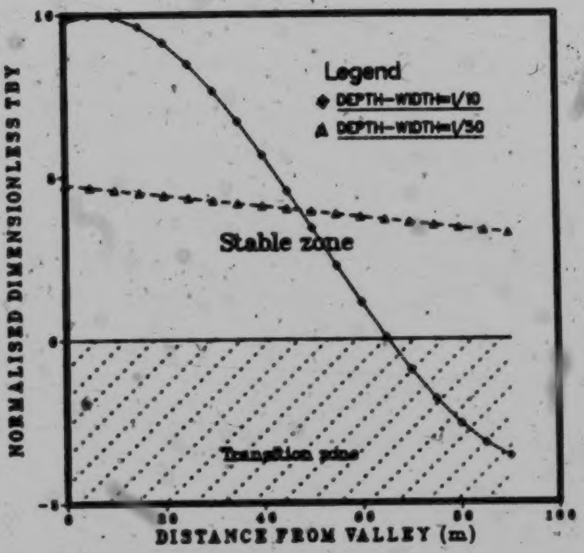


Figure 66. Effect of basin depth on basin stability

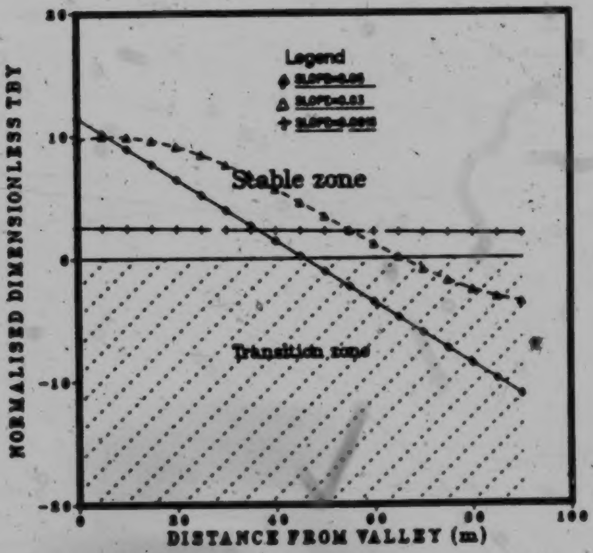


Figure 67. Effect of regional slope on basin stability.

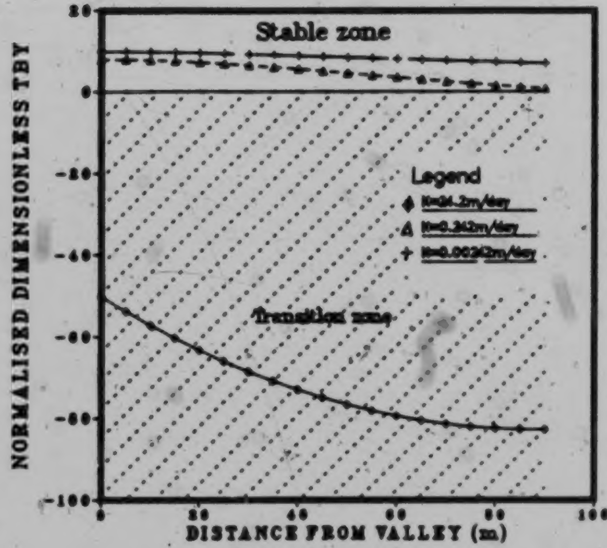


Figure 68. Effect of hydraulic conductivity on basin stability.

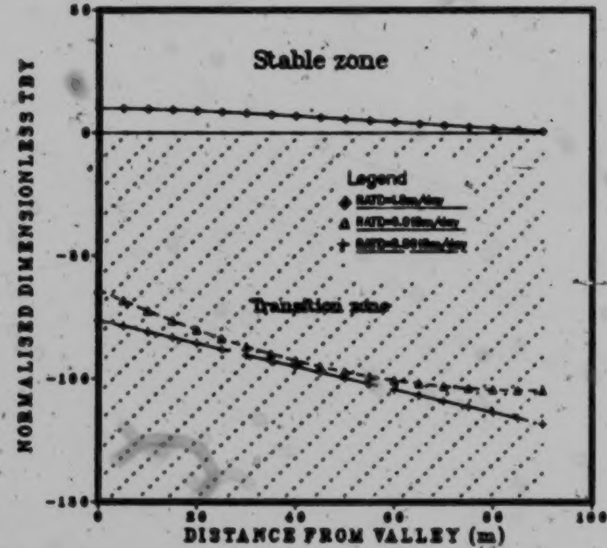


Figure 69. Effect of the ratio of precipitation to hydraulic conductivity on basin stability.

#### 4.5.3.2.9 Summary

It is possible to analyze the stability range of a basin under development by estimating the cumulative flow (TBY) through the system during the transient period of development. The stable range is marked by positive TBY; the boundary between stable and unstable conditions by low positive, low negative, or zero TBY and a steady water table; and instability is recorded by high negative TBY and a water table that declines rapidly with time.

A basin may be developed in the discharge area to obtain stable conditions, whereas, the same basin under the same initial and boundary conditions, and having the same hydrogeologic environmental parameters may result in groundwater mining if development takes place in the recharge area.

In particular, the tendency to obtain stable conditions in response to development increases with: 1. increasing precipitation; 2. decreasing permissible drawdowns; 3. decreasing well depths; 4. increasing depth to water table; 5. decreasing basin depth; 6. decreasing regional slope; 7. increasing precipitation to hydraulic conductivity ratios.

# 5. GROUNDWATER MANAGEMENT IN ROSS CREEK BASIN

## 5.1 General statement

Thus far, hydrological considerations and a methodology have been presented for planning the optimal exploitation of regional groundwater resources in the Plains Regions of Alberta. In particular, an understanding of the interrelationship between the hydrogeologic environment and variations thereof, and the resulting groundwater regime has been used in the determination of the best withdrawal sites while considering the stability of the regional setting simultaneously. The resulting guidelines may be employed in various ways in evaluating the groundwater resources of different Plains Regions environments.

In this chapter, the methodology is applied in the management of groundwater in Ross Creek Basin in order to demonstrate one of the possible ways in which the method is applicable.

## 5.2 The problem

Ross Creek Basin is situated in southeastern Alberta, in Improvement District No. 1. It covers an area of approximately 608 km² southeast of the City of Medicine Hat, bounded on the north by the Ross Creek valley at the Trans-Canada Highway, and on the south by the Cypress Hills divide (Figure 70). In terms of the Dominion land-survey system used commonly in the central and western provinces of Canada, the area is located in townships 8, 9, 10, 11 and 12, ranges 2, 3, 4 and 5, west of the 4th meridian. The northern two-third of Ross Creek Basin is in the driest portions of the province of Alberta. Strictly speaking, precipitation is low, evapotranspiration balance is little for the often dry years of this area (Ophori and Tóth, 1983).

Despite the poor conditions of water supply, plants and animals still abound in this area, that require water for their daily subsistence. Water is required mainly for domestic, irrigation, and municipal purposes. In spite of the dry nature of Ross Creek Basin, several springs discharge groundwater perennially at various topographic elevations and at various rates. The reliability of the springs is not known, but their existence indicates a continuous supply of water to the groundwater system. Neither is the effect of spring development on the entire groundwater system known, thus, Alberta Environment is unwilling to issue licenses for their development as requested by the local inhabitants. The degree to which the entire

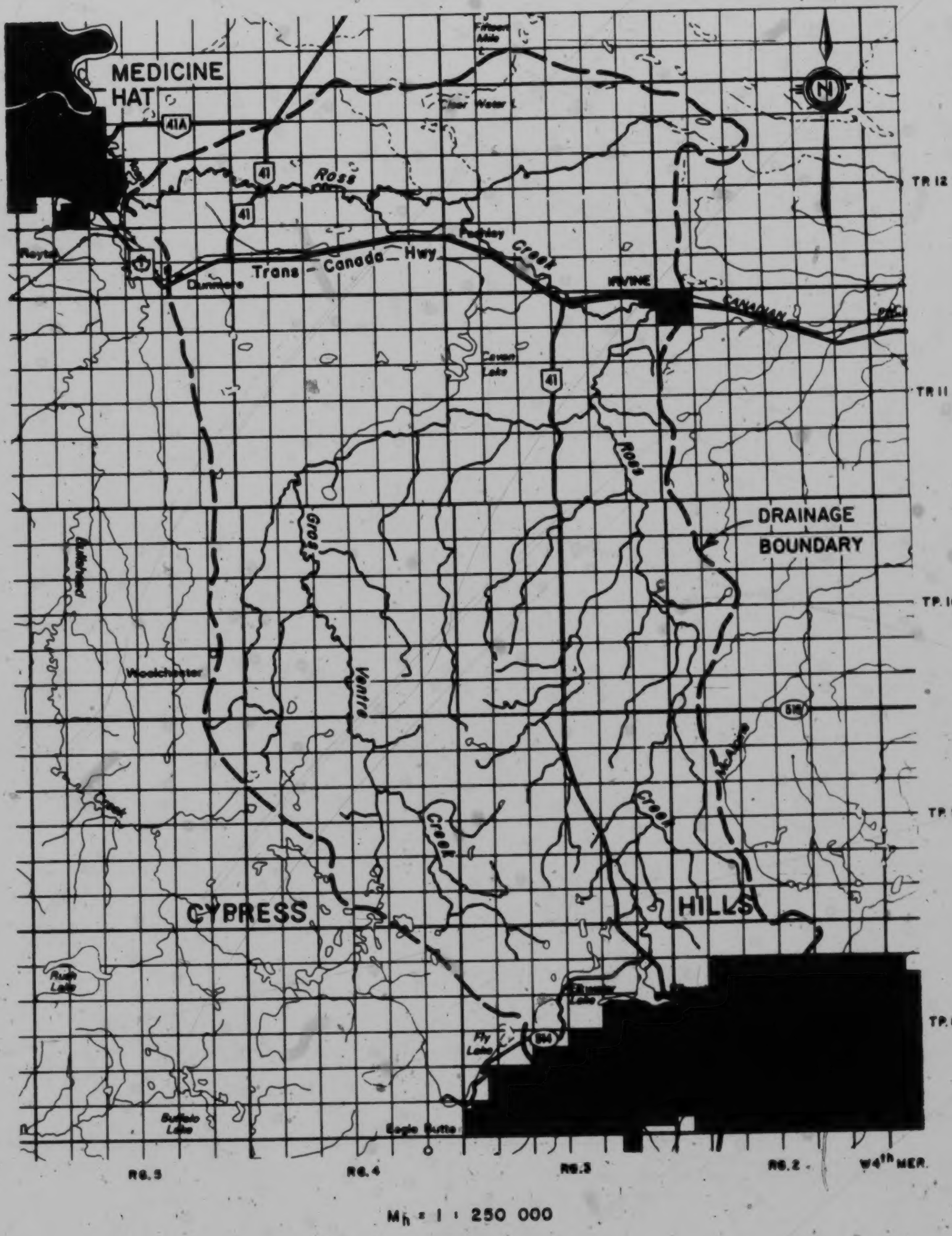

MEDICINE HAT
41A
41
Ross
Fifteen Mile
Clear Water L.
Royal
1
Dunmore
Trans - Canada - Hwy
Creek
IRVINE
CANADIAN
PACI
Cavan Lake
Ross
Gros
DRAINAGE BOUNDARY
Bullshead
Woolchester
Ventre
Creek
Creek
Creek
515
CYPRESS
HILLS
Rush Lake
Fly Lake
514
Buffalo Lake
Eagle Butte
N
TR 12
TR 11
TP. 10
TP. 9
TP. 8
RG. 5
RG. 4
RG. 3
RG. 2
W4th MER.
Mh = 1 : 250 000

Figure 70. Location of Ross Creek Basin.

groundwater system is interconnected determines its response to development at any particular site. Only when this degree of interconnection is known can a planned exploitation of the groundwater be made.

The problem at hand, therefore, is to evaluate the groundwater regime, that is: the occurrence, movement, quality, and possible use of the groundwater, and to provide a model for the management of the available groundwater resources in the basin.

### 5.3 The hydrogeologic environment and aspects of management

The determination of the groundwater regime requires that the hydrogeologic environment of the basin be known. Thus, field studies were carried out with the major objective of ascertaining the hydrogeologic environment, and to obtain a few aspects of the groundwater regime. The hydrogeologic environment was determined in a similar manner as outlined in chapter 2, but actual field measurements were emphasized. The most salient features of the environment are summarized below from the report by Ophori and Tóth (1986) which is included in Appendix 3 for ease of reference.

Cross sections across the basin (Appendix 3, Figure 2) reveal a topography whose slope is dominated in one direction (north-south direction). The east-west component of groundwater flow is, therefore, assumed to be negligible regionally compared to the north-south component; that is, a two-dimensional cross-sectional flow is assumed. Although the basin consists of different formations geologically, an analysis based on hydrogeologic units defines two different, but hydraulically uniform sets of units: two aquifers and an aquitard. One of the aquifers is restricted to a little zone around the lower Ross Creek valley (Figure 71). The second aquifer lies in the property of the Cypress Hills Provincial Park. An analysis of water level fluctuation in observation wells produced a recharge from precipitation to the groundwater system of 113 mm or $3.0\times10^{-4}$ m/day. On the basis of topography, hydrogeologic units and climate, a hydrogeologic environment is defined for Ross Creek Basin as illustrated in Figure 71. Figure 71 lies between the ATR and BTR categories of Plains Regions environments.

Mapping of surficial manifestations of groundwater movement reveals a groundwater regime of various orders of flow systems (local, intermediate and regional). Whereas springs and other discharge features mark the discharge ends of the flow systems, and the conspicuous absence of these features the recharge areas, a relatively large area has been mapped as

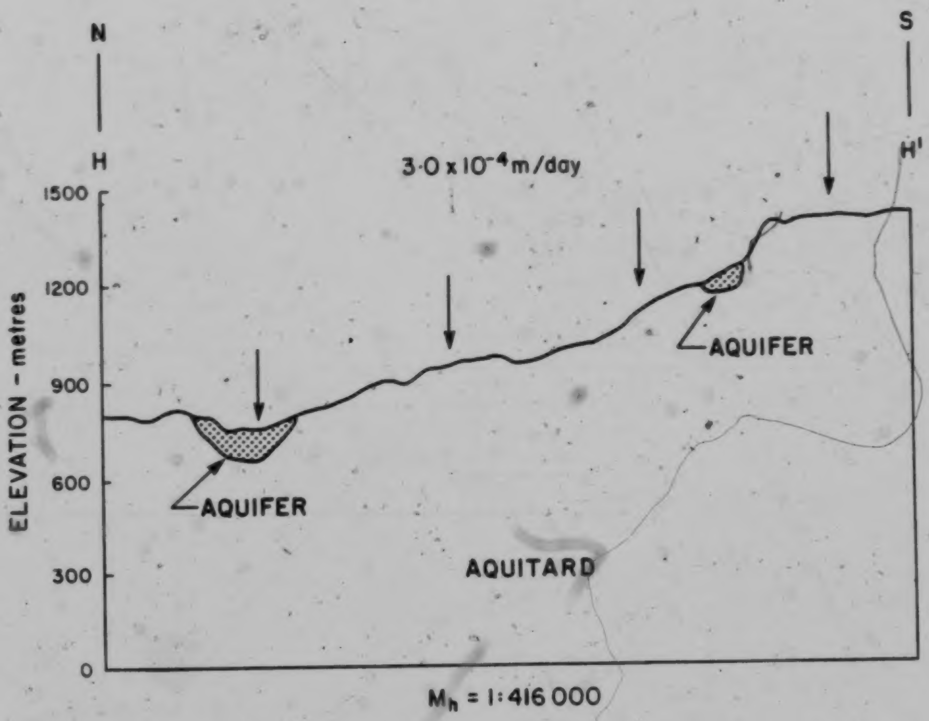
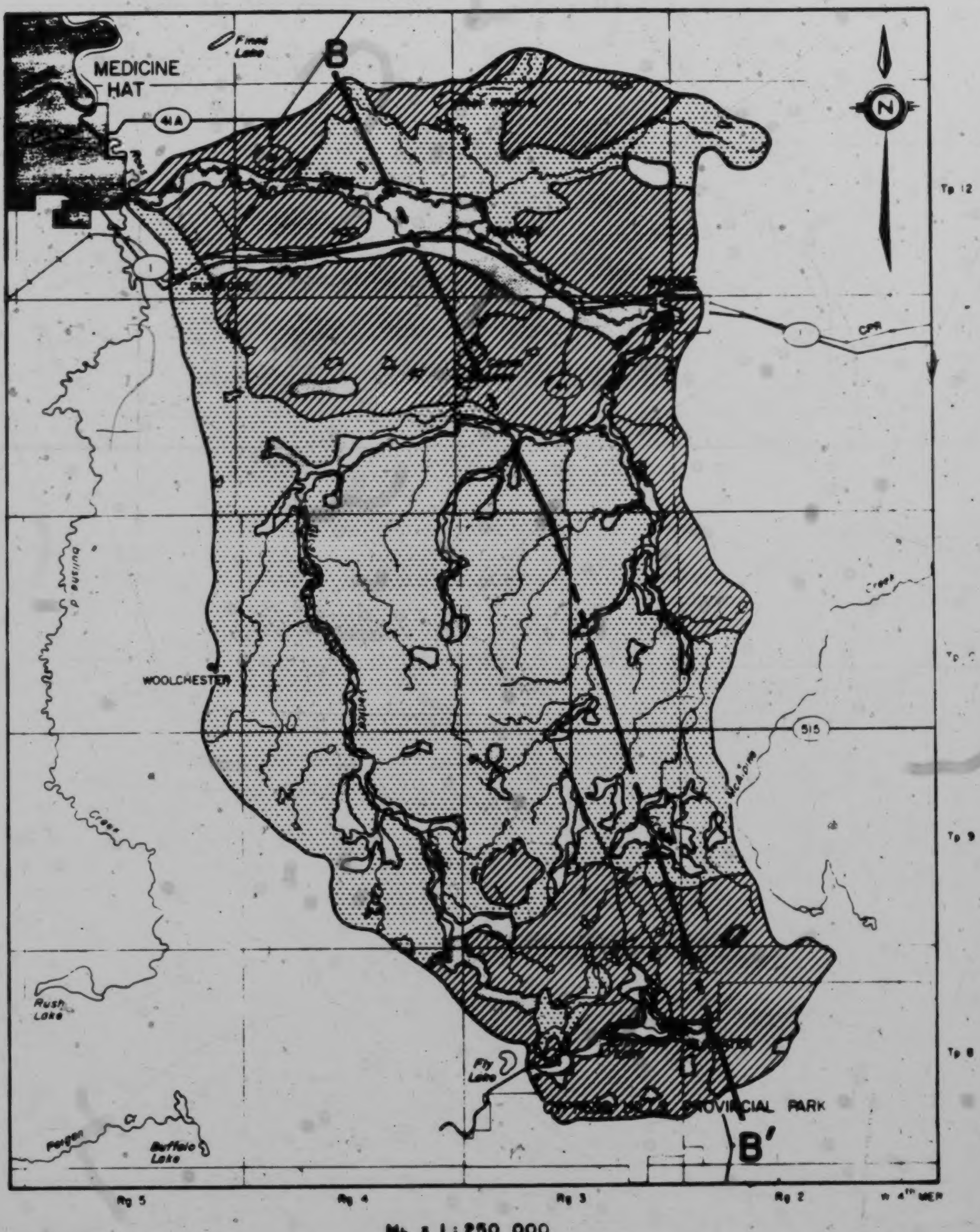

N
H
S
H'
$3{\cdot}0 \times 10^{-4}$ m/day
1500
1200
900
600
300
0
ELEVATION - metres
AQUIFER
AQUIFER
AQUITARD
$M_h$ = 1:416 000

Figure 71. Hydrogeologic environment of Ross Creek Basin.

uncertain on the basis of the features (Figure 72). Aquifer hydraulics shows the lack of reliable aquifers in the basin, and indicates that less than $4.1 \times 10^{-3}$ $m^3/s$ of groundwater can be produced from a few localities only, to cause complete mining of the resource in twenty years. A safe production rate of $1.8 \times 10^{-3}$ $m^3/s/km^2$ that would cause no permanent water level lowering was calculated from the analysis of dynamic reserves. In addition to the good agreement between the groundwater chemistry and the identified flow systems, up to 50 percent of the groundwater is potable, within treatable limits.

Several issues call for groundwater management in Ross Creek Basin: 1. groundwater is scarce not because there is no recharge from precipitation, but because there is a lack in appropriate storage and transmission materials (aquifers) from which it can be produced as required; this is evident from the two very different groundwater productivity scenarios predicted on the basis of aquifer hydraulics on one hand, and dynamic reserves calculations on the other; 2. as a consequence of the first point above, several springs are discharging, evaporating and wasting groundwater, and there is the fear to issue licenses for their use because the groundwater regime is not understood; 3. runoff along Ross Creek and its tributaries is substantial relative to demands for water in the basin, but it is flashy in nature and requires storage for the time of need; and 4. annual, seasonal, and daily fluctuations in hydrometeorological studies of the basin are all in favour of a water management plan.

Under these conditions, the best use that can be made of the subsurface resources is their conjunctive operation with surface water resources. However, a prerequisite to such an operation is the presence of adequate aquifers that are capable of being recharged artificially or naturally through induced infiltration. The lack of such aquifers, according to our present knowledge of the basin, precludes a consideration of a conjunctive operation. Nevertheless, in order to make the best use of the available groundwater resources in the basin, a logical basis for the operation of the springs in the basin is formulated. The formulation is based on an understanding of the groundwater regime and its response to possible development.


MEDICINE HAT
41A
B
B'
1
CPR
515
WOOLCHESTER
Rush Lake
Fly Lake
Buffalo Lake
Creek
Creek
CYPRESS HILLS PROVINCIAL PARK
Tp 12
Tp 9
Tp 8
Rg 5
Rg 4
Rg 3
Rg 2
W 4th MER

Mh = 1 : 250 000

LEGEND

▨ Recharge area
☐ Discharge area
░ Uncertain area

B-B' TRACE OF SECTION IN FIGURE 10

Figure 72. Distribution of recharge and discharge areas.

SOURCE-Adapted From Ophori & Toth, 1983

## 5.4 Theoretical considerations

### 5.4.1 Steady-state groundwater regime

To gain a more detailed understanding of the groundwater movement through the basin than permitted by field studies, the construction of a flow systems map was necessary. Unfortunately, the construction of meaningful hydraulic cross sections from field measurements of water levels was impossible because of the lack of observation points that are deep enough for this purpose. However, a theoretical flow pattern was determined for the basin with the aid of field observed data.

Figure 73 shows the mathematical model representation of Ross Creek Basin in two dimensions. As discussed previously, the section whose trace is indicated in Figure 70 was chosen carefully so as to approximate a direction perpendicular to the regional trend, and parallel to the regional hydraulic gradient. It is presumed that the flow pattern in this cross section would be representative of the regional groundwater flow.

Essentially, the domain is made up of the hydrogeologic environment (Figure 71). The right vertical boundary which coincided with the crest of the Cypress Hills was assumed to be impermeable. Also, a bottom impermeable boundary was located at the contact of the Alberta Group and Milk River Formation. For the steady-state runs, the top boundary was the water table, which was assumed to coincide with the surface topography because water level data were not enough to define it in the required detail. The justification for this assumption is that water level data indicated the water table to lie within 20 m from the surface in all cases. Furthermore, this assumption is documented elsewhere (Jones, 1960; Farvolden, 1961; Tóth, 1962). The left vertical boundary was assumed to be impermeable in some cases, and as a constant head boundary in others. Equation (8), together with the domain and boundary conditions described above completed the boundary value problem that was solved.

The hydraulic head distribution was determined using two different finite element numerical models. First the model of Frind (1971) as described in section 3.3 was employed, and the top boundary was treated by specifying a hydraulic head value equal to the surface elevation at each node along the boundary. Several runs were processed varying the left boundary conditions and the hydraulic parameters (Kx and Ky) of the units comprising the aquifers and aquitard, in order to simulate measured hydraulic heads at the points indicated on

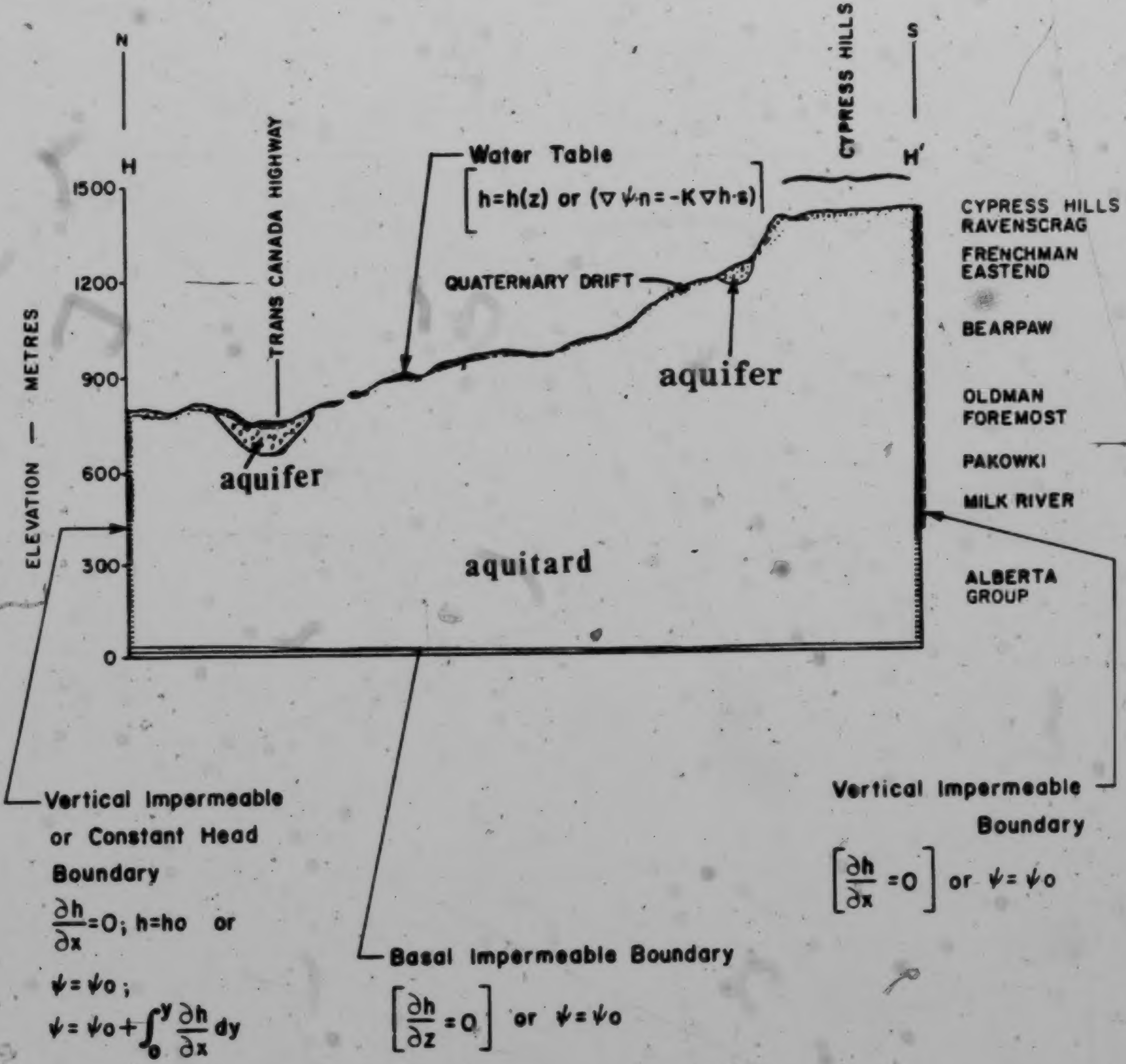

N
S
H
H'
CYPRESS HILLS
TRANS CANADA HIGHWAY
Water Table
[h=h(z) or (∇ψ·n=-K∇h·s)]
QUATERNARY DRIFT
aquifer
aquifer
aquitard
1500
1200
900
600
300
0
ELEVATION — METRES
CYPRESS HILLS
RAVENSCRAG
FRENCHMAN
EASTEND
BEARPAW
OLDMAN
FOREMOST
PAKOWKI
MILK RIVER
ALBERTA
GROUP
Vertical Impermeable or Constant Head Boundary
∂h/∂x=0; h=ho or
ψ=ψo;
ψ=ψo+∫₀ʸ ∂h/∂x dy
Basal Impermeable Boundary
[∂h/∂z=0] or ψ=ψo
Vertical Impermeable Boundary
[∂h/∂x=0] or ψ=ψo

Figure 7.3. Two-dimensional mathematical model of Ross Creek Basin.

Figure 74. Initially, field estimated hydraulic conductivities were used, and homogeneous and isotropic conditions were assumed. The aquifers were maintained at a hydraulic conductivity of two orders of magnitude higher than that of the aquitard throughout the simulations. Stepwise adjustments in hydraulic conductivities were made with reference to values obtained from literature. After each simulation, simulated hydraulic heads were compared with the measured point hydraulic heads. Changes were then made as necessary and simulations repeated. Measured point hydraulic heads were used for the calibration because no reliable hydraulic cross section could be constructed from the available water level data. A certain degree of doubt may be associated with results obtained in this manner, but it is felt that a range of flow patterns resulting from this analysis would be reliable. Best match between observed and simulated hydraulic heads was produced with the hydraulic conductivity values indicated in Table 11. The resulting hydraulic head distribution is displayed in Figure 74. The correlation between observed and simulated hydraulic heads is better at the lower topographic elevations than it is at higher elevations, probably due to the better coincidence of the surface topography with the water table at the lower elevations. Nevertheless, it is felt that the results of the mathematical model are in good correlation with the general nature of the observed hydraulic head distribution in the basin.

Having determined the hydraulic head distribution and regional hydraulic conductivities, it was necessary to estimate the flow pattern of groundwater. Also, the flow pattern would serve a second role of further calibration of the model. For this purpose, the mathematical model was reformulated in terms of stream functions as described in section 3.3. Equation (19) was then solved for Ross Creek Basin with the appropriate boundary conditions to generate a flow pattern map (Figure 75) directly. The scale of Figure 75 has been exaggerated vertically but a direct interpretation of the flow pattern is possible.

Qualitatively, groundwater flow in the basin consists of local, intermediate and regional flow systems. Figure 75 shows that about 40 percent of the groundwater in the basin is recharged in the areas around the Cypress Hills. One-half of this water is discharged through a local flow system at a lower topographic elevation adjacent to the hills, and through an intermediate system into the Gros Ventre Creek valley, whereas the remaining one-half is discharged at the Ross Creek valley through a regional flow system. Figure 75 also indicates that the remaining 60 percent of the groundwater moving through the basin is recharged

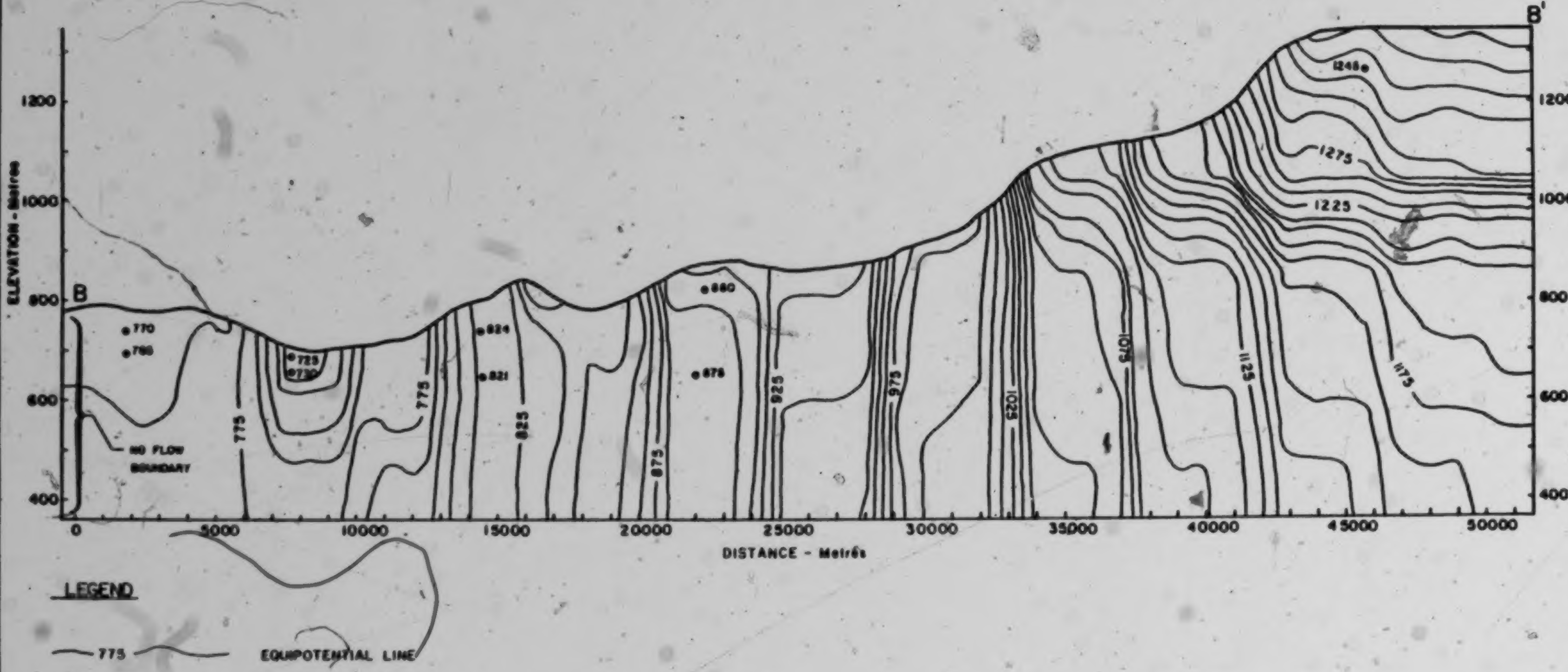

B
B'
ELEVATION - Metres
1200
1000
800
600
400
0
5000
10000
15000
20000
25000
30000
35000
40000
45000
50000
DISTANCE - Metres
770
760
723
730
824
821
880
875
1246
775
775
825
875
925
975
1025
1075
1125
1175
1225
1275
NO FLOW BOUNDARY
LEGEND
775
EQUIPOTENTIAL LINE

Figure 74. Map showing simulated hydraulic head distribution.

Table 11. Adjusted hydraulic conductivities used in numerical models.

| Formation | | Hydraulic conductivity (m/s) | |
|---|---|---|---|
| Aquifer | Aquitard | Horizontal (Kh) | Vertical (Kv) |
| | Cypress Hills | $8.4 \times 10^{-5}$ | $8.4 \times 10^{-7}$ |
| | Ravenscrag | $4.4 \times 10^{-5}$ | $4.4 \times 10^{-7}$ |
| | Frenchman | $4.5 \times 10^{-6}$ | $4.5 \times 10^{-8}$ |
| | Eastend | $4.5 \times 10^{-5}$ | $4.5 \times 10^{-7}$ |
| | Bearpaw | $4.2 \times 10^{-6}$ | $4.2 \times 10^{-8}$ |
| | Oldman/Foremost | $4.4 \times 10^{-5}$ | $4.4 \times 10^{-7}$ |
| | Pakowki | $4.2 \times 10^{-6}$ | $4.2 \times 10^{-8}$ |
| | Milk River | $8.4 \times 10^{-5}$ | $8.4 \times 10^{-7}$ |
| | Quaternary Drift | $2.1 \times 10^{-6}$ | $2.1 \times 10^{-8}$ |
| Quaternary Drift | | $4.2 \times 10^{-3}$ | $4.2 \times 10^{-5}$ |

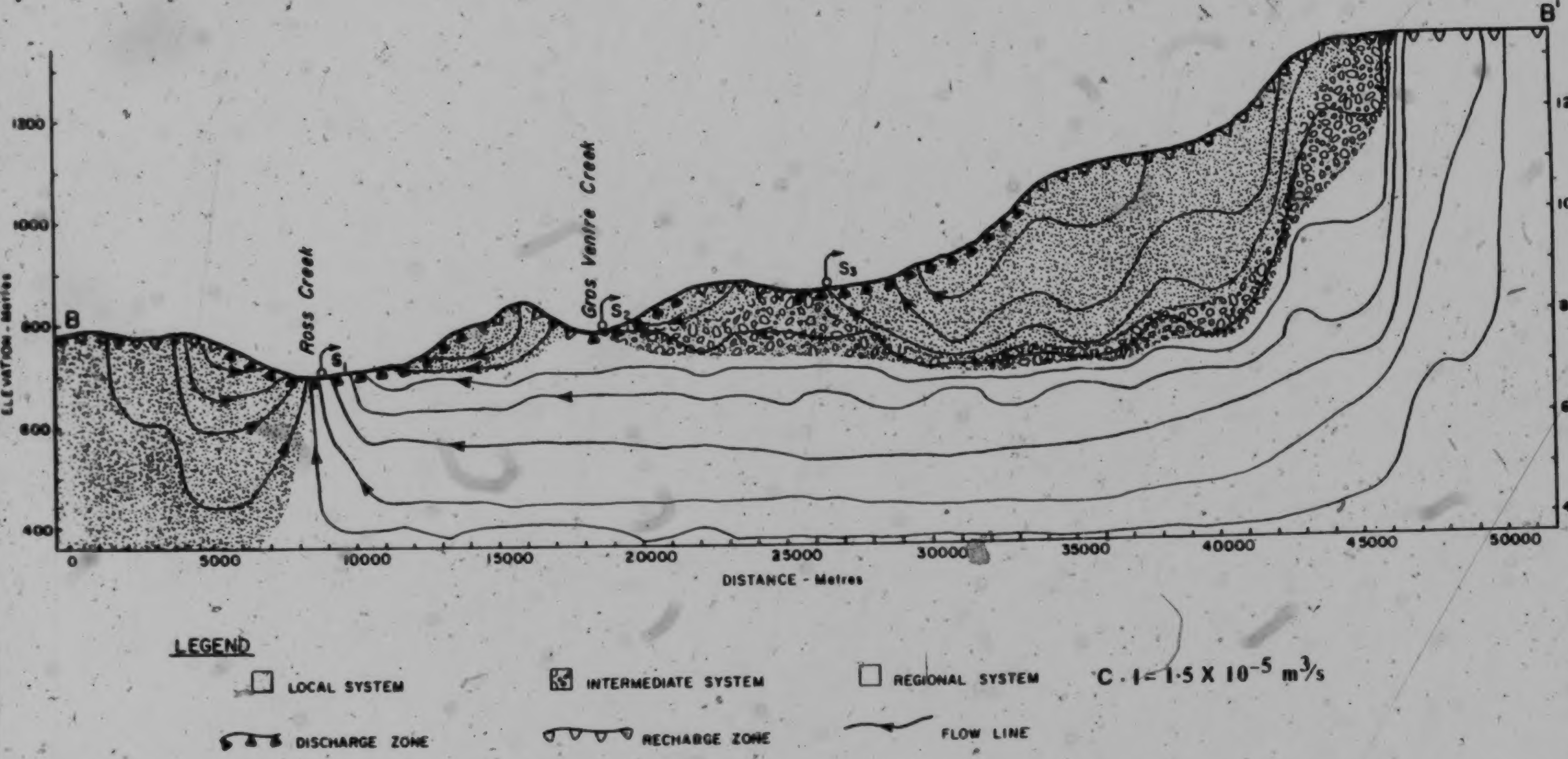

B
B'
Ross Creek
Gros Ventre Creek
S1
S2
S3
ELEVATION - Metres
400
600
800
1000
1200
0
5000
10000
15000
20000
25000
30000
35000
40000
45000
50000
DISTANCE - Metres
LEGEND
LOCAL SYSTEM
INTERMEDIATE SYSTEM
REGIONAL SYSTEM
DISCHARGE ZONE
RECHARGE ZONE
FLOW LINE
C · I = 1·5 X 10⁻⁵ m³/s

Figure 75. Map showing simulated flow pattern.

through local topographic highs on adjacent sides of Gros Ventre and Ross Creeks.

The calculated recharge-discharge picture of the flow pattern in the basin leads to a series of alternating recharge and discharge areas (Figure 75). A comparison of Figures 72 and 75 shows a good correlation between recharge and discharge areas of both figures, giving credit to the surficial mapping technique for groundwater reconnaissance studies in Ross Creek Basin. Furthermore, the coincidence of recharge and discharge areas is a further indication that the flow pattern, and thus, the hydraulic head distribution determined by the models are representative of the basin's groundwater regime.

Quantitatively, Figure 75 can be analyzed easily to determine the total discharge through the system. Recognizing that the contour interval of Figure 75 is $1.5 \times 10^{-5}$ $m^3/s$, a total discharge (equal to recharge) of $2.0 \times 10^{-4}$ $m^3/s$ is estimated for the basin. This is the natural basin yield (Freeze, 1968; Freeze and Witherspoon, 1968) or the dynamic reserves (Tóth, 1982), the lower limit of groundwater that can be developed theoretically from Ross Creek Basin on sustained basis. The calculated basin yield converts to 118 mm or 22 percent of the total precipitation of 531 mm. The similarity between the value of dynamic reserves obtained by this theoretical method and the value (113 mm) estimated from measured water level fluctuations is remarkable. However, if one recognizes that it is the same parameter being calculated by two different methods, their similarity could be expected.

The estimated dynamic reserves are probably higher than expected. Most casual observers would agree that the prospects for significant amount of recharge in the basin is poor because of the sparse cover of drought-resistant grasses and shrubs, and mean annual actual evapotranspiration rates that approach or exceed mean annual precipitation (Laycock, 1967). In fact, Wong (1974) estimated a water surplus of zero for most of the northern parts of the basin, indicating that groundwater may not occur at all. Whereas it is possible that the former approach may have overestimated the groundwater recharge because the basin does not appear to be sufficiently wet, the water balance method appears to underestimate groundwater recharge for the following reasons: 1. the continuous discharge from springs in the basin must be maintained by a source of water; 2. fluctuations in water levels in wells of the area indicate a source of supply; 3. the water table is sub-parallel to the surface topography, a condition that may not exist in the absence of groundwater recharge; 4. phreatophyte patches are found in several zones in the basin, and a perennial tributary to Ross Creek has been formed from discharge by springs at one site (SW-14-12-4W4); 5. a water surplus of up to 152 mm has been estimated at the Cypress Hills Plateaux (Wong, 1974), the area that is believed to be the main zone of recharge of water into the basin in the present study; and 6. groundwater recharge values of up to 20 percent of annual precipitation have been observed in New Mexico (Stephens, 1985), in areas where mean annual evapotranspiration is several times greater than mean annual precipitation. A determination of the actual quantitative relationship between rates of groundwater discharge and the features above is beyond the scope of this thesis. Until such a relationship is quantifiable, the disparity in the estimates of groundwater recharge by these methods cannot be reconciled.

The model by Frind (1971) was employed in the calibration process because it is cheap to execute, and a large number of runs was required to achieve a satisfactory calibration. However, this model is not equipped to handle the necessary transient simulations. To achieve this objective, all the calibrated parameters of the groundwater system were passed on to the model by Davis and Neuman (1983). This model as described earlier (Section 4.3) was executed once to redetermine the steady-state hydraulic head distribution. During this run, a pressure head value of zero was specified at each node along the surface topography to simulate the water table. The final steady-state hydraulic head distribution obtained by this model was similar to that of the previous model, with the largest discrepancy between nodal values being less than 1 m. Once it was obtained, the natural steady-state distribution of hydraulic heads served as the initial conditions for the transient analysis of the basin's groundwater regime that follows.

### 5.4.2 Response of the groundwater regime to development

The purpose of the transient analysis was dual: 1. to determine the best possible well location, its rate of production, and its role in meeting water demands in the basin; and 2. to determine the ultimate steady-state groundwater flow pattern and thus, advice on an appropriate management strategy.

To achieve this purpose, the transient model described in section 4.3 was used to develop the basin's groundwater system. The initial conditions were those of the natural steady-state hydraulic head distribution (Figure 74). Since the hydrogeologic environment (Figure 71) resembles the ATR and BTR environments (Figure 22) of the Plains Regions in which the aquifer lies in the discharge area, the best position of a well in terms of maximum rates of production is easily identifiable. Only the aquifer at the lower Ross Creek valley is developed here because the other aquifer lies in the property of the Cypress Hills Provincial Park. According to Figure 58, such a well should be located at the middle of the aquifer. With the well located thus (point W, Figure 76), the model was run to steady state several times with various possible conditions of recharge from precipitation and available drawdown.

During the first run (run 1) a recharge rate of $3.0\times10^{-4}$ m/day, distributed evenly along the water table was allowed into the system to simulate the recharge rates estimated both theoretically and through dynamic reserves calculations. A maximum permissible drawdown equal to two-third of the total available drawdown was assumed. To simulate pumping in the well, the water level in the well was held constant at a hydraulic head value equivalent to the permissible drawdown throughout the run. The model was then run until steady state was reached.

The interpretation of the results from the runs was based mainly on the earlier findings that: 1. the stability of a groundwater system under development is marked by positive values of transitional-basin-yield (TBY) and a steady water table; 2. the transition between stable and unstable conditions is associated with low negative, zero, or low positive TBY and a steady water table; and 3. instability is registered by high negative TBY and an unsteady water table.

Results of this run showed that although steady state was achieved eventually, the calculated TBY (Table 12) is negative indicating that the system is in the transition state, an unwanted situation. The results also showed that groundwater movement in the vicinities of springs $S_2$ and $S_3$ (observed during the hydrogeological mapping) remained upward at the end of the run. The water table still intersects the land surface in these areas, thus, groundwater is discharged continuously by $S_2$ and $S_3$. It is inferred from these results that, whereas enough water is recharged into the system to maintain groundwater discharge at $S_2$ and $S_3$, the hydraulic conductivity of the aquitard is lower than one needed to conduct enough water through the aquitard to the aquifer and thus to the well, in order to ensure the continuous stability of the system.

The situation is different in the area around $S_1$. The natural discharge area around $S_1$ is converted to a recharge area by pumping from the well. This is another detrimental effect of such a pumping scheme, as downstream water users will be deprived of their share of the

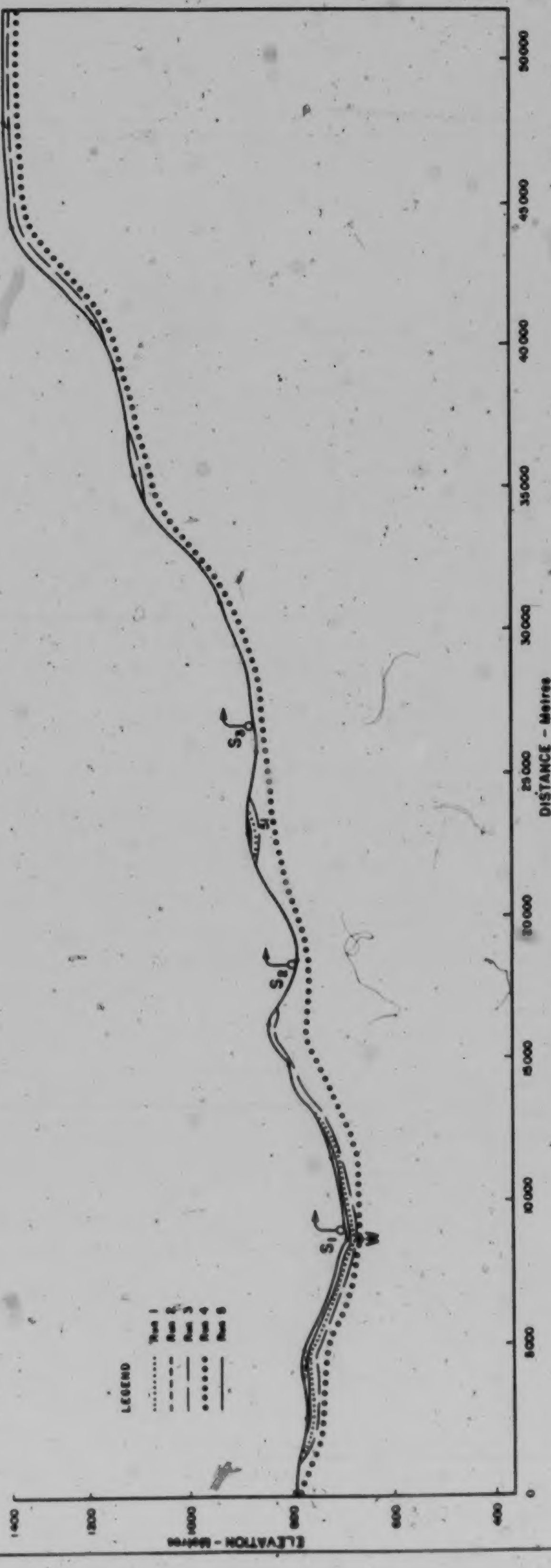


Figure 76. Position of water table for runs 1 to 5.

Table 12. TBY values for runs 1 to 5.

| Run | TBY ($m^3$) |
|---|---|
| 1 | -459 |
| 2 | -1 244 |
| 3 | -3 122 |
| 4 | -2 740* |
| 5 | 96 |

* [illegible]dy state attained

natural supply of water along the creek.

The second run (run 2) was designed to represent a situation in which the recharge into the system is areally restricted; thereby simulating the entry of water into the system through recharge areas only, depression focussed recharge, and so on. During this run, the same recharge rate as in run 1 was concentrated at the nodes in the recharge areas along the water table, all other parameters being the same as in the first run. The results and interpretations thereof were similar in every respect to those in the preceding run, except that a little more negative TBY was calculated (Table 12). This is an indication that the system will be less stable to development if the recharge were areally restricted.

In run 3, all the parameters of run 1 were used except that the recharge rate was reduced by one-half to $1.5\text{x}10^{-4}$ m/day in order to simulate the conditions of a drier year. Once more, results similar to those of run 1 were obtained with an appreciably more negative TBY. In addition to the system being much less stable, an interesting deduction is that $S_2$ and $S_3$ continue to discharge groundwater. This situation suggests that if the actual groundwater recharge is half of the calculated value, most of the springs at the intermediate and high levels in Ross Creek Basin will continue to flow in the presence of development at the lower Ross Creek valley.

Run 4 simulated the conditions of a completely dry year, therefore, recharge into the system was zero, and all other parameters remained as they were in run 1. No steady state was obtained. Drawdown was slow but steady, and the run was stopped after enough time was allowed to obtain the required results. In this case, the water table fell below the ground level everywhere in the system (Figure 76). Thus, the development of the basin during a long period of no recharge into the system will cause unstable conditions, accompanied by the drying up of the springs.

Following these findings, several runs were processed to determine the optimal pumping plan for the basin. For each consecutive run, the permissible drawdown was reduced systematically and the value of TBY was examined until it became positive. In all these runs, the calculated recharge rate of $3.0\text{x}10^{-4}$ m/day was used, and all other parameters corresponded to those of run 1. A positive value was first calculated in a run (run 5) in which the permissible drawdown was approximately equal to one-third of the available drawdown.

In run 5 which represents stable conditions of development in the basin, permissible drawdown is one-third of the available drawdown, springs $S_2$ and $S_3$ are discharging groundwater, and there is no discharge from $S_1$. This is viewed as the ultimate groundwater regime that will result from the optimal development of groundwater in Ross Creek Basin.

Another result obtained from the model calculations is the rate of pumping at the production well which corresponds to the optimal development. From run 5, the calculated pumping rate (obtained directly as model output) is $8.8\text{x}10^{-5}$ $m^3/s$. Field investigations indicate presently that the aquifer stretches for a length of approximately 1 000 m perpendicular to the modelled cross section. If this width is considered, together with a safety factor of 0.5, a total pumping rate of $4.4\text{x}10^{-2}$ $m^3$ /s is obtained. This value may be considered as a first approximation of the groundwater yield at the most favourable site from the basin hydrological view point. It should be noted that this groundwater yield is based on a two-dimensional analysis, and may or may not correspond exactly to a three-dimensionally derived yield.

Considering these theoretical results, the following deductions can be made about groundwater management in Ross Creek Basin: 1. the only aquifer of significance that was detected during the field studies in the basin has a potential production rate of $4.4\text{x}10^{-2}$ $m^3/s$; this rate is greater than the demand for total municipal supply ($4.8\text{x}10^{-3}$ $m^3/s$), and less than the demand for domestic ($5.8\text{x}10^{-2}$ $m^3/s$) and irrigation ($2.6\text{x}10^{-1}$ $m^3/s$) uses; 2. at the estimated safe rate of production, springs at intermediate and higher elevations will continue to discharge groundwater, thus, most of these springs can be licensed for use at their natural rates of discharge, without danger to the groundwater regime; and 3. at the calculated safe pumping rate, however, natural discharge into the lower Ross Creek is captured completely, thereby jeopardizing the water rights of downstream water users; the tradeoff between the advantages and disadvantages of this effect was not considered further in this study.

In order to satisfy the water rights of downstream users, that is, to allow some natural discharge into Ross Creek at the production site, a number of runs was processed with continuous reduction in the permissible drawdown. It was observed that pumping becomes negligible for this objective to be satisfied.

### 5.5 Summary

The regional hydraulic gradient in Ross Creek Basin is dominated by the north-south component, thus, the basin's groundwater flow can be analyzed in a two-dimensional vertical cross section. Aquifer hydraulics shows that $4.1\text{x}10^{-3}$ $m^3/s$ of groundwater can be pumped from a restricted locality (aquifer) in order to mine the groundwater in twenty years. A greater portion of the basin is an aquitard. Water level fluctuation and numerical simulation analyses indicate that $3.0\text{x}10^{-4}$ m/day of water flows through the basin. The components above constituted the hydrogeologic environment, that is, the framework with which the groundwater regime was analyzed.

If the basin's groundwater is developed so that two-third of the available drawdown is used up at the best possible site of development, transitional-basin-yield (TBY) is negative, indicating the tendency toward instability. To maintain stable conditions during development, only one-third of the available drawdown may be utilized to produce $4.4\text{x}10^{-2}$ $m^3/s$ of groundwater or 13 percent of the total water demands ($3.3\text{x}10^{-1}$ $m^3/s$). During such a development, the springs at intermediate to high topographic elevations will discharge groundwater continuously, thus, such springs may be licensed for local use. However, the springs at the lower Ross Creek valley will be obliterated by this development and deprive downstream users of their water supply from the creek. Groundwater pumpage may have to be reduced to zero approximately, in order to ensure water supply to downstream users. Therefore, groundwater pumpage is not a significant source of water to a centrally operated water system in the basin.

# 6. SUMMARY

A hydrological characterization of the Plains Regions of Alberta in the form of a baseflow analysis of 32 selected basins was carried out in order to gain a basic understanding of the regional distribution, movement, and discharge of groundwater. Plots of specific baseflow rate versus basin area [q(A)], display a regularity. According to the regularity, with increasing area of basin at initial low values of basin area, specific baseflow rate increases rapidly. At and above a certain area of basin (approximately 6 000 km$^2$), the rate of increase of specific baseflow rate decreases with further increase in basin area, thus specific baseflow rate tends toward a constant value.

The existence of a baseflow regularity reveals several factors pertinent to regional groundwater flow in the Plains Regions, as portrayed by regional groundwater discharge to streams: 1. regional groundwater discharge is neither haphazard nor random, it occurs in a systematic manner; 2. all other factors being equal, the size of the basin appears to control groundwater discharge; 3. in basins of 6 000 km$^2$ and larger, conditions of groundwater discharge become relatively constant, therefore, such basins may be considered as "type units" or "type environments" for regional groundwater studies in this region. It is justifiable to assume that all the water moving into the type environment is discharged from it, that is, local, intermediate and regional flow systems as the case may be, have been developed completely.

A hypothesis was advanced to explain the observed regularity. In general, in the Plains Regions of Alberta, and at the large scale at which the baseflow analysis was carried out, as one moves from high order streams through lower order streams to the streams of lowest order, one transects sub-basins of increasing area from high to low topographic elevation. The sub-basins are often bounded by small regularly occurring topographic divides with their streams occupying their lower central portions. The water table often coincides with the surface topography, and generates flow systems of different order. Consequently, water recharged into the system at the highest divide is lost through a local flow system to the adjacent stream, and through intermediate and regional systems to streams that occur further downslope. The resultant effect is that streams at high elevations and those at low elevations receive less and more water respectively, than is recharged through their associated divides. This unequal sharing takes place in a systematic manner that produces a baseflow regularity.

The concepts of the hydrogeologic environment and groundwater regime were reviewed as an attempt to develop a comprehensive list of the factors which control the hydrogeologic setting of drainage basins in the Plains. The factors are all related to the topography, geology, and climate of the setting. The hypothesis above was verified through numerical-model simulations of two-dimensional vertical sections of typical Plains Regions' drainage basins. It is only through such models that the effects of the complicated hydrogeologic environment on regional groundwater distribution, movement and discharge can be understood, and be usefully employed in regional groundwater development and management.

The models confirm for the first time, the regular distribution of steady-state regional groundwater discharge and its relation to basin area in the Plains. Quantitatively, the regularity (regularity type 1) indicates the occurrence of greater discharge rates in regional discharge areas (areas of low topography) than in the regional recharge areas (areas of high topography). In this regard, and recognizing that calculated specific discharge equals specific basin yield, the regularity may guide in the exploration for, and exploitation of groundwater in the regional sense. During groundwater exploitation in basins that are smaller than the type environment, care must be exercised in selecting the basin with the optimum specific basin yield. On the other hand, in basins of the same size as the type environment and larger, such a selection is not critical since every basin would discharge groundwater at about the same rate.

An inverse regularity, a mirror image of regularity type 1 is obtained if the smallest basin is considered to occur at the lowest topographic elevation, and the basin area is increased in the direction of increasing elevation. A comparison of regularity type 1 and its inverse gives another clear indication that a greater specific basin yield is associated with a basin at a low elevation (regional discharge area and including local recharge regions) relative to that of a basin of the same size at a high elevation (regional recharge area and including local discharge regions) in a regional setting.

Several factors of the hydrogeologic environment affect the simple relationships demonstrated by regularity type 1 to different degrees: viz, the ratio (n) of the depth to the horizontal impermeable boundary to the width of the main basin, the average regional slope, the width of the main basin, the width of the smallest sub-basin, and the size and position of aquifers within the system.

A reduction of n from 0.5 to 0.05 to simulate shallow Plains Regions environments changes regularity type 1 to regularity type 2, a straight horizontal line indicating a constant specific discharge for the entire range of basin area. The shallow nature of the environment does not allow the development of intermediate and regional flow systems, thus, no interbasinal transfer of water occurs. Each sub-basin has an isolated flow system that is unconnected to adjacent systems. As a tool for exploration, regularity type 2 indicates that specific basin yield is constant for every sub-basin. In other words, groundwater can be produced from every sub-basin at the same rate, and every sub-basin is a type environment in a shallow setting of the Plains Regions. A more detailed analysis of the effect of n shows that n rather than the sizes of the sub-basins is the sole agent controlling the interchange between regularity types 1 and 2.

The effect of an increase in the regional slope of the setting is to enhance specific discharge in basins at lower topographic elevations relative to basins at higher elevations. The enhancement may be so great that regularity type 1 is transformed to regularity type 2 which in turn may be transformed to regularity type 3, a linear discharge curve. This means that the isolated local flow systems of shallow environments are converted to include intermediate and regional flow systems with increase in slope. A combination of an adequate depth of a basin and a high regional slope on the other hand causes a linear distribution of specific basin yield. In this case (regularity type 3), no two sub-basins possess similar specific basin yield, thus, there is no type environment for the setting. Each sub-basin must be evaluated in isolation for groundwater resources. The main basin has a specific basin yield that is much greater than that of the smallest sub-basin at high elevation. All other sub-basins possess linearly increasing intermediate basin yields.

In the presence of depths of basins that are adequate for the development of local, intermediate and regional flow systems, such systems are better developed with increasing size of basin.

The size of the smallest and incremental sub-basin produces no significant effect on the flow pattern in a regional setting, every other parameter being equal.

An areally extensive aquifer in an environment acts as a conduit which tends to collect and transmit water from high to low basins. The shallower the position of the aquifer in the system, the more the effect of water transfer is felt. Also, an increase in the ratio of aquifer to aquitard hydraulic conductivity increases the transfer of water from high to low elevations. In a shallow environment, the occurrence of an aquifer alone might be the only condition for the development of intermediate and regional flow systems to unite the otherwise isolated local flow systems. In such cases, the groundwater discharge conditions of the setting will be characterized by regularity type 1 as opposed to the anticipated regularity type 2.

The effect of a partial aquifer in the environment is similar to that of the areally extensive aquifer but, in addition, discharge in sub-basins at the up-slope end of the aquifer is depleted preferentially relative to that of the sub-basins down-slope of the aquifer.

No new discharge curves are formed other than those developed under homogeneous conditions by the occurrence of aquifers in the environment. Neither is the regular distribution of groundwater discharge destroyed by the presence of aquifers, although flow patterns are interconvertible. Thus, in the regional sense, the Plains Regions environments behave as to mask the effects of inhomogeneity to a low level.

A knowledge of the discharge regularity type in an environment implies a knowledge of the controlling environmental parameters, and hence the possible flow pattern, whereas, a knowledge of the environmental parameters may be used to infer the possible regularity, and then the possible flow pattern. That more groundwater exists in the discharge area than in the recharge area of a regional setting under virgin conditions can be quantified through the consideration of regularities in discharge, therefore, relative to the average water balance of the setting, excess and deficit of water should attend discharge and recharge areas, respectively.

Numerical-model simulations of unit basins in the Plains Regions were run to define the principles governing the relation between well yield and well location, and between well location and basin stability during development, for basin planning and management purposes. The numerical models which approximate the groundwater flow equation for variably saturated porous media in vertical sections were used to develop various categories of the Plains Regions environment systematically.

The models are the first to show the detailed relations between transitional-basin-yield (TBY), sustainable-basin-yield (SBY), and the location of the well in the basin. In general, TBY is highest in the discharge area, and it decreases gradually toward the recharge area. On the other hand, SBY increases from the discharge area to the recharge area. From a basin hydrological viewpoint, these are key relations to planning the optimal location of development sites in a basin.

The positioning of the site of development within a basin is a major factor in determining the proportion of precipitation that can be captured into the basin. A pumping well in the recharge area of a basin induces more infiltration from surface water into the basin, and captures less of the natural discharge from the system, relative to a well in the discharge area. Well locations for maximum and minimum capture from surface water do not necessarily coincide with the divide and valley of the system. The locations are controlled by the basin geometry as well as the rate of precipitation.

As a result of the natural flow regimes in a groundwater basin, more water is available for use if development takes place in the discharge area relative to the recharge area. Therefore, in a basin with uniform hydrogeologic environmental parameters, exploration for groundwater should be initiated in the discharge area, and moved only gradually toward the recharge area for compelling reasons.

The optimal development of a basin's water resources is controlled to a large extent by the use for which the water is intended; optimal well location depends upon whether the use requires a high or low volume of supply, high or low rate of supply, or both. Under conditions of sufficient precipitation, use types 1 and 3 are favoured by the location of a withdrawal site in the discharge area of a basin, whereas, use types 2 and 4 require that the site of development be moved toward the recharge area.

A combination of the TBY and SBY versus distance curves produces a composite curve, the tool for optimizing well location for some specified uses. The maximum and minimum points of this curve correspond to the best locations of wells intended to serve use types 3 and 4 respectively.

Various factors of the hydrogeologic environment influence well location in the regional setting, and determine the level of planning required to locate the wells optimally. Use type 1 is synonymous with maximum SBY, use type 2 with maximum TBY, use type 3 with maximum SBY and TBY, and use type 4 with minimum SBY and TBY.

For optimal development, use types 1, 2, and 3 should be located in the recharge, discharge, and midline areas of a basin respectively. Use type 4 is favoured most often at the divide. The presence of a partial aquifer in the basin causes an exception to this general rule. The midpoint of the aquifer is the best location for use types 1 and 3, whereas, use types 2 and

4 should be sited at the valley and divide respectively, under these conditions.

All other controlling parameters being equal, higher SBY and/or TBY may be obtained if wells are sited: in basins with higher rainfalls; at deeper levels in the basin; in basins with higher regional slopes for SBY, and in the discharge areas of basins with higher regional slopes for TBY; in deeper basins; in basins with deeper water tables; in materials with higher hydraulic conductivities for SBY, and with lower hydraulic conductivities for TBY; in extensive aquifers SBY, and in restricted aquifers for TBY; in aquifers in discharge area rather than in aquifers in recharge areas.

Through the modeling approach presented herein, it has been possible to analyze the stability of a basin under development. This thesis demonstrates for the first time that under conditions of restricted rainfall, a basin can be developed on a sustained yield basis by locating a well in the discharge area, whereas groundwater mining results in the same basin under the same conditions, but with the well in the recharge area..

The tendency to obtain stable conditions in response to development increases with: increasing precipitation; decreasing permissible drawdowns; decreasing well depths; increasing depth to water table; decreasing basin depth; decreasing regional slope; increasing precipitation to hydraulic conductivity ratios.

One significant finding is that, the existence of a high hydraulic conductivity alone is not a sufficient proof that the basin is a reliable source of groundwater supply. This statement is borne out by the fact that in the presence of precipitation rates that are significantly low in comparison to the hydraulic conductivity, groundwater mining occurs earlier than when the hydraulic conductivity is lower. It is the ratio of precipitation rate to hydraulic conductivity that is more important in basin hydrology.

In conducting field studies designed to manage groundwater in the Plains Regions, care should be exercised in defining hydrogeologic environmental parameters such as precipitation rates, regional slopes, basin depth, hydraulic conductivities of geologic materials, and sizes and positions of aquifers within the region. It is especially important to know the midpoint, and the down- and up-slope ends of high permeability lenses.

Where applicable, management decisions such as well depth, permissible drawdown and so on, should be based on existing field conditions rather than on rules that have been determined empirically and engraved in literature.

The modeling approach and findings in this thesis were applied in the management of groundwater in Ross Creek Basin, Alberta. This approach is particularly suited to the problem of groundwater management in the basin not only because the possible rate of production can be predicted, but because other components of the groundwater system can be considered simultaneously.

The classical method of aquifer hydraulics predicted a yield of $4.1 \times 10^{-3}$ m$^3$/s to mine the available groundwater in the basin completely in twenty years. The calculation is based mainly on values of transmissivity. Transmissivity is only a measure of the ability of geologic materials to transmit water, therefore, it provides no information about the availability or source of the water to be transmitted through the system. Recognizing that transmissivity can be calculated empirically as the product of hydraulic conductivity and the thickness of the unit, and recollecting one of the earlier conclusions about hydraulic conductivity, high transmissivities do not necessarily signify good production potentials from the point of view of basin hydrology. It is the ratio of the hydraulic conductivity to precipitation rate that is more important in basin hydrology, thus, yields estimated without a consideration of precipitation rates may be in error. Often, aquifers pumped at the rates predicted through aquifer hydraulics are not mined as expected in twenty years. Another disadvantage in the application of this method in Ross Creek Basin is that the response of the entire groundwater system to possible development is required to propose a licensing procedure for the springs of the area. Aquifer hydraulics is incapable of providing this information.

Application of the basin hydrological models showed that groundwater can be produced at a sustainable rate of $4.4 \times 10^{-2}$ m$^3$/s at the most favourable site in Ross Creek Basin if the interest of downstream users is not represented in the management plans. However, if downstream users are considered, possible groundwater pumpage is negligible with respect to the uses for which it is needed. The basin's groundwater system is to be aided artificially to enhance natural discharge from the numerous springs. In cases where discharge from springs does not contribute to the water supply of downstream users, licenses may be issued to the local inhabitants for the use of such springs.

The summary will not be complete without a mention of limitations that might be imposed by several assumptions in the study, the most obvious of which are outlined as follows: Firstly, all the models presented are in two-dimensional vertical cross sections implying that the third dimension is infinite. This assumption may not be absolutely correct when applied to a real groundwater basin because the actual dimensionality of the problem is not considered accurately. However, several conclusions based on one- and two-dimensional models have been found to hold in their corresponding three-dimensional models (Freeze, 1969c, 1971; Winter, 1976, 1978). The two-dimensional cross sections consist of the three fundamentally different groundwater flow regimes (recharge, transfer, and discharge regimes) of major importance in this study, and are therefore adequate for the present considerations. Secondly, the transient simulations used the steady-state gravity flow of groundwater as their initial conditions, and the end results were those of steady state established as a result of development. Inherently, it has been assumed that the system responds to development at every spatial location in a similar manner during the intervening transient period. At present, there is no reason to believe that the system behaves otherwise. Thirdly, the quantity and quality of data required to analyze the systems in their true nature are not readily available. Simplifying assumptions and subtitutions had to be made in some cases. However, this situation is not perculiar to this study; it is often the case in most groundwater studies of the scale considered herein.

Despite these limitations, if groundwater is to be managed in units as large as drainage basins, the missing link between aquifer hydraulics and basin hydrology must be found. This thesis is another step further in that direction.

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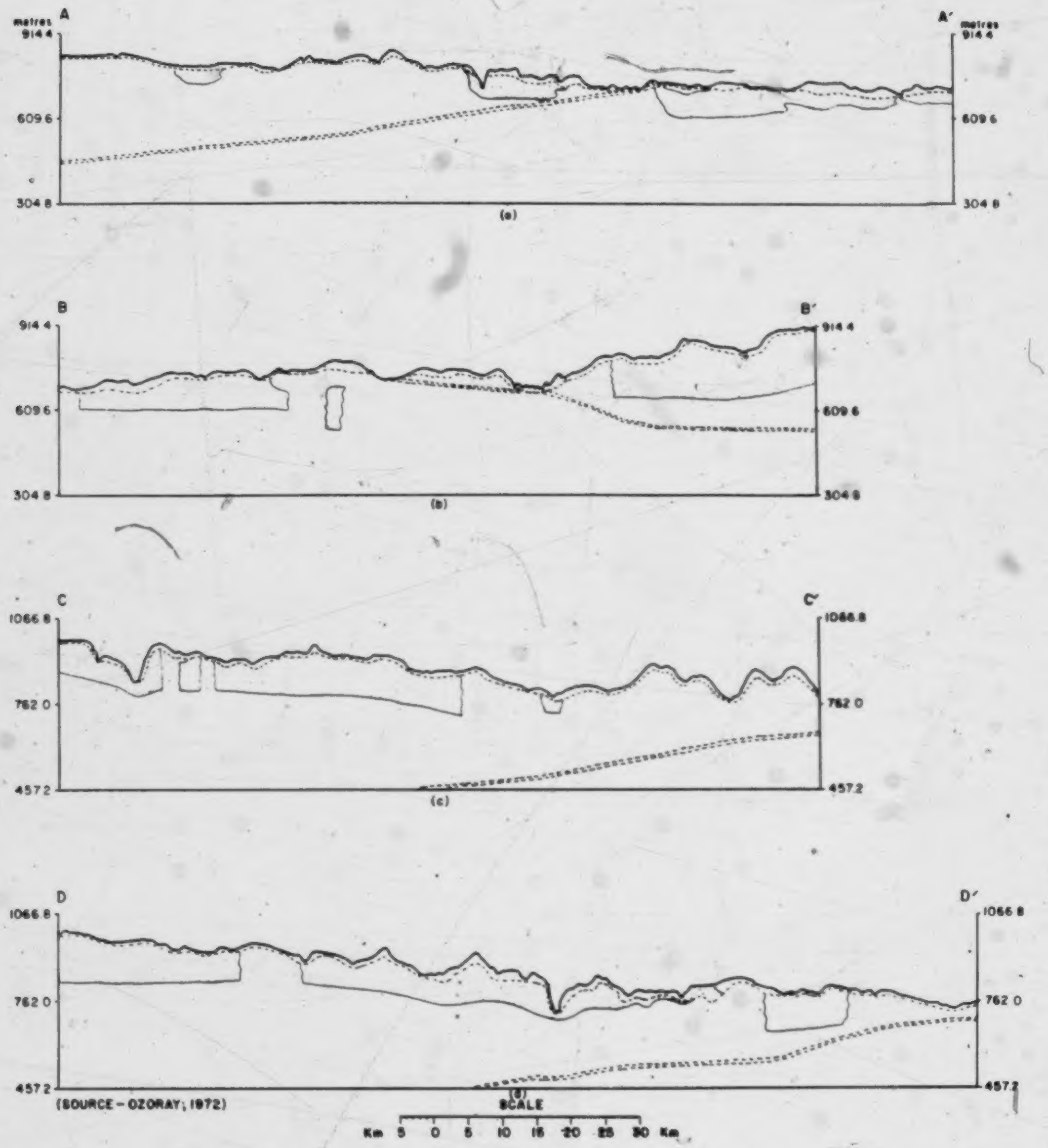
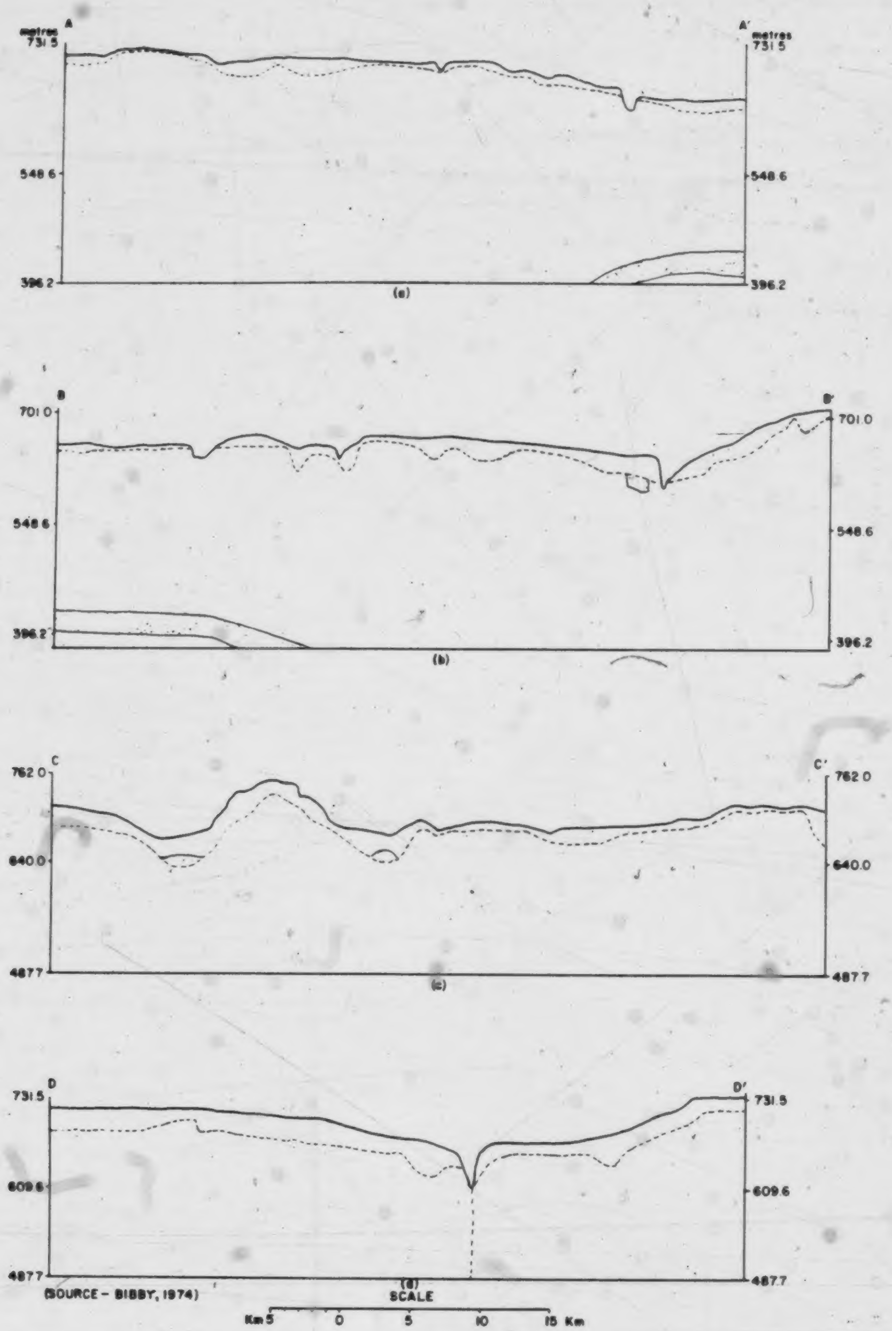
# Appendix 1 - Profiles of topography, geologic units, and hydrogeologic units in the Plains Regions of Alberta

| | |
|---|---|
| ——— | BOUNDARY OF HYDROGEOLOGIC UNITS |
| - - - - - - - - | BOUNDARY OF GEOLOGIC UNITS |
| ▨ | AQUIFER $(Q \geq 2{\cdot}0 \times 10^{-3}\ m^3/s)$ |
| ☐ | AQUITARD $(Q < 2{\cdot}0 \times 10^{-3}\ m^3/s)$ |


A
A′
metres
914.4
609.6
304.8
(a)
B
B′
914.4
609.6
304.8
(b)
C
C′
1066.8
762.0
457.2
(c)
D
D′
1066.8
762.0
457.2
(d)
SCALE
Km 5 0 5 10 15 20 25 30 Km

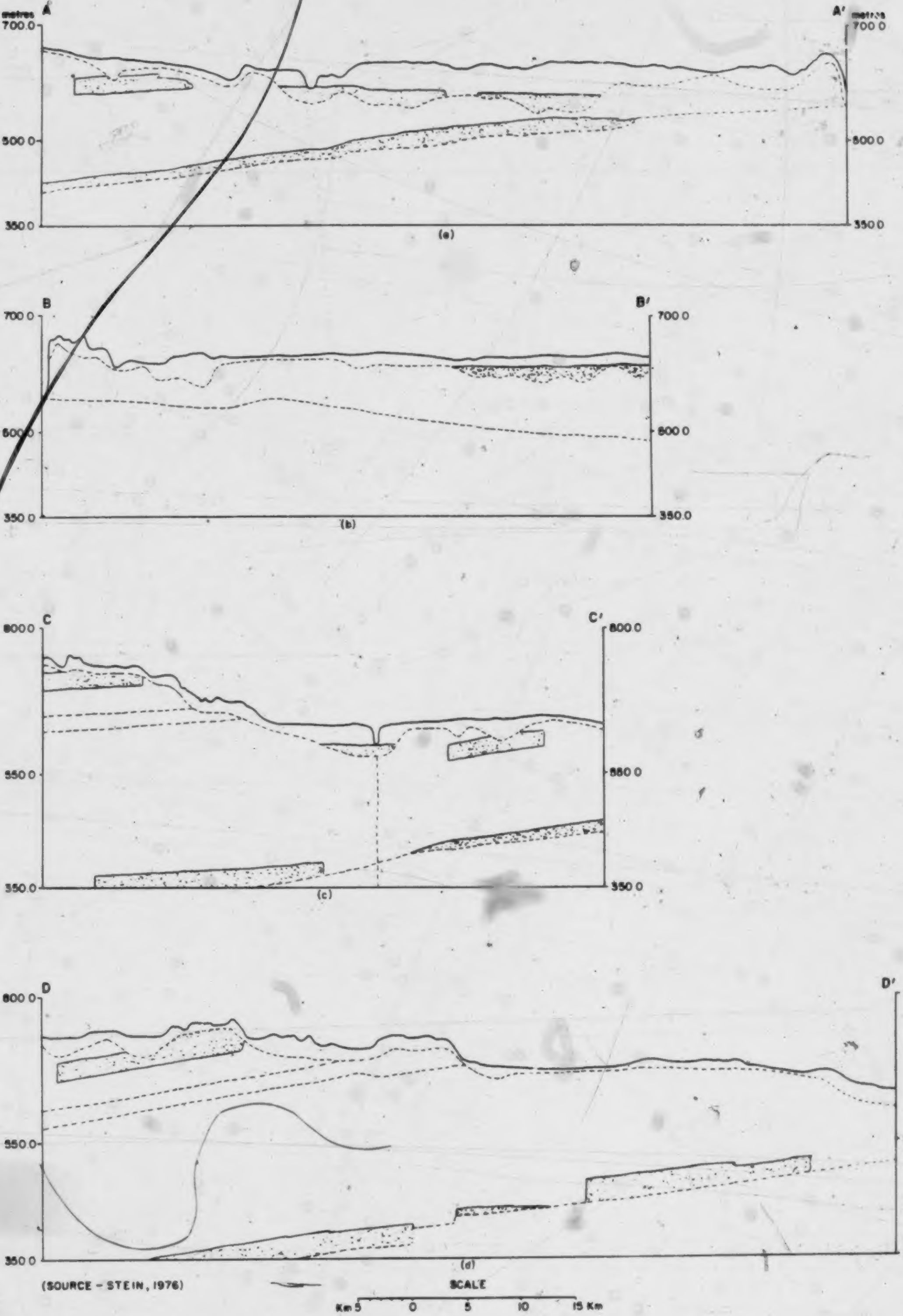
(SOURCE – OZORAY; 1972)

Plate 1



(SOURCE – BIBBY, 1974)

Plate 2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metres
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700.0
500.0
600.0
350.0
350.0
(a)
B
B'
700.0
700.0
600.0
600.0
350.0
350.0
(b)
C
C'
800.0
800.0
550.0
550.0
350.0
350.0
(c)
D
D'
800.0
800.0
550.0
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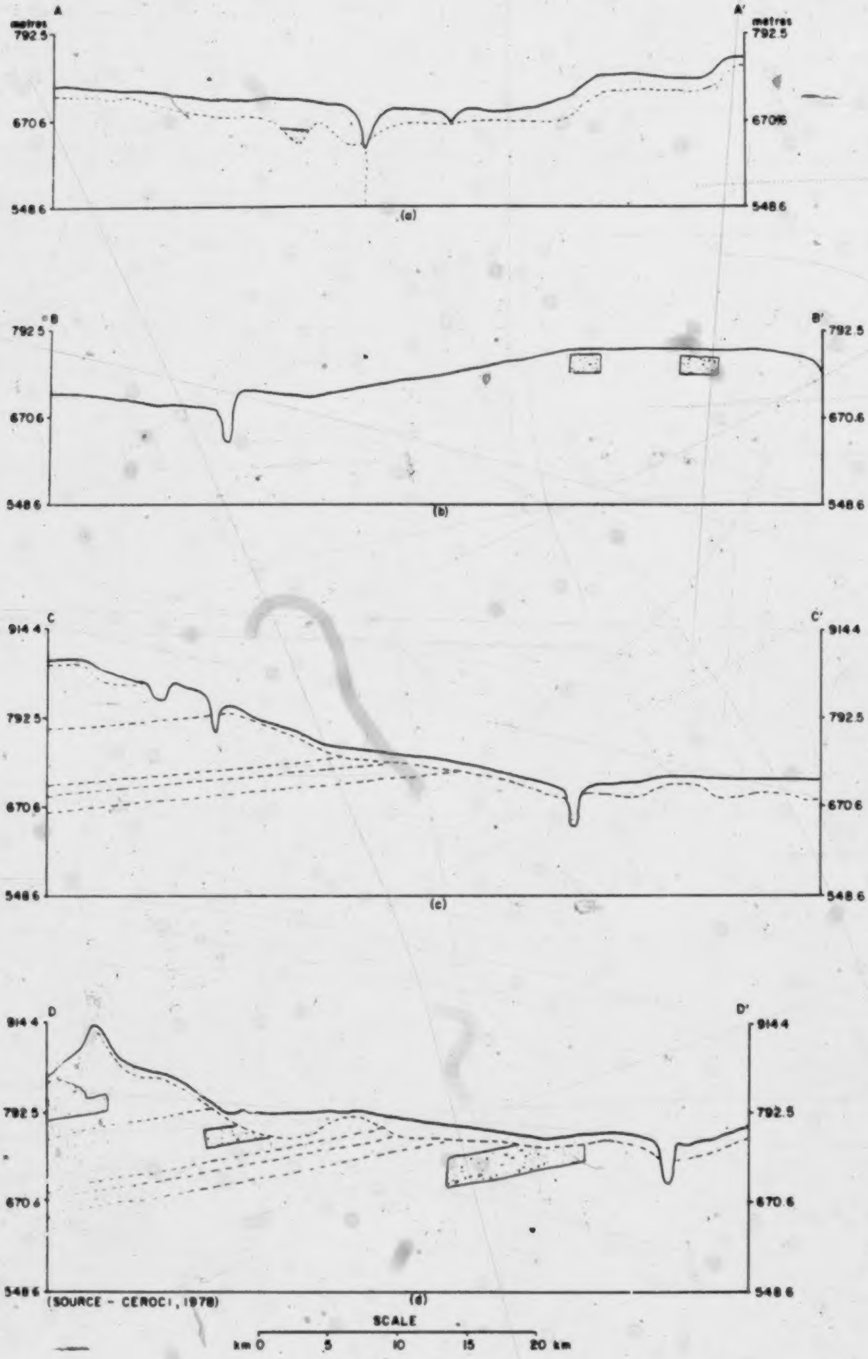
350.0
350.0
(d)
SCALE
Km 5
0
5
10
15 Km

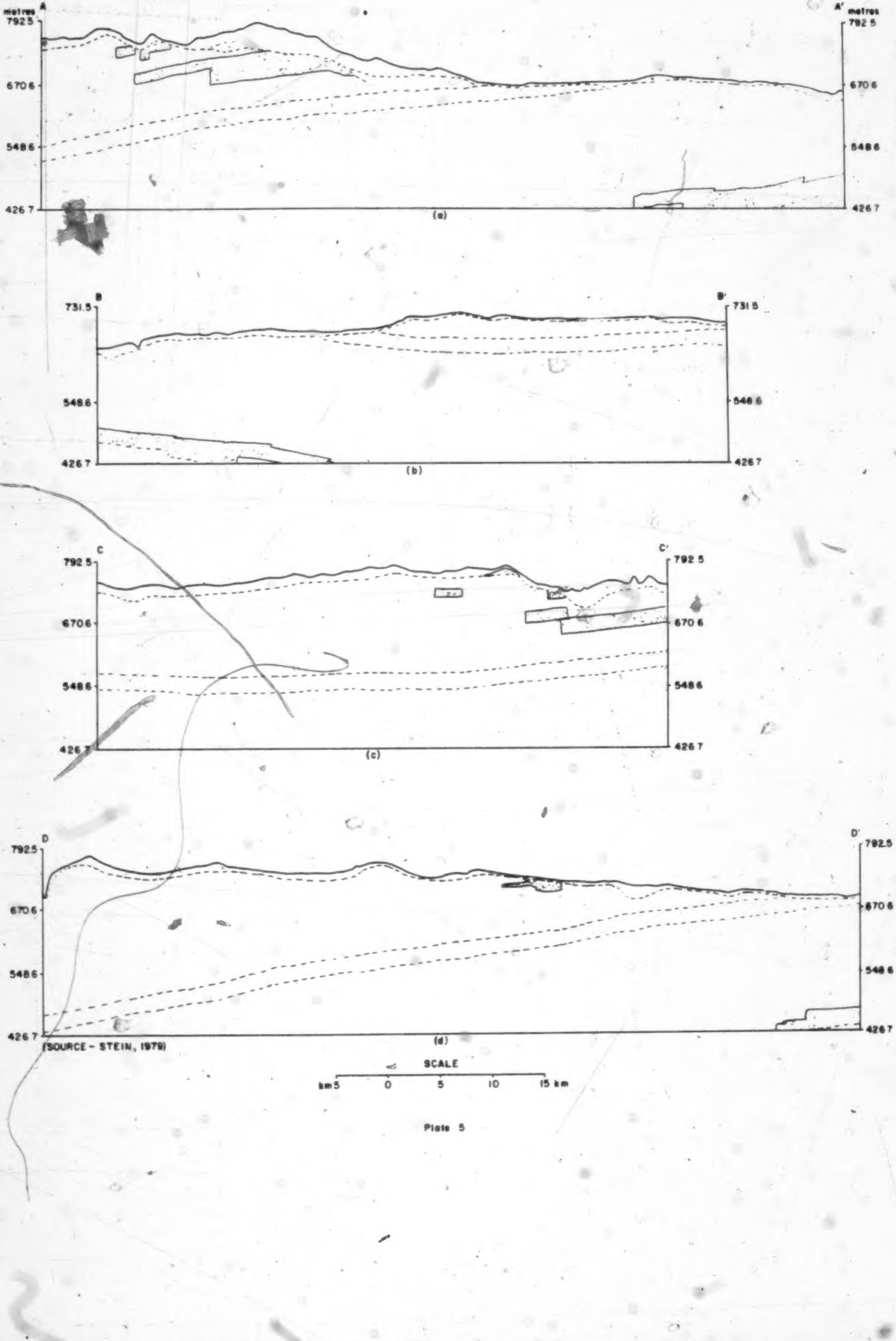
(SOURCE - STEIN, 1976)

Plate 3



(SOURCE - CEROCI, 1978)

Plate 4


metres
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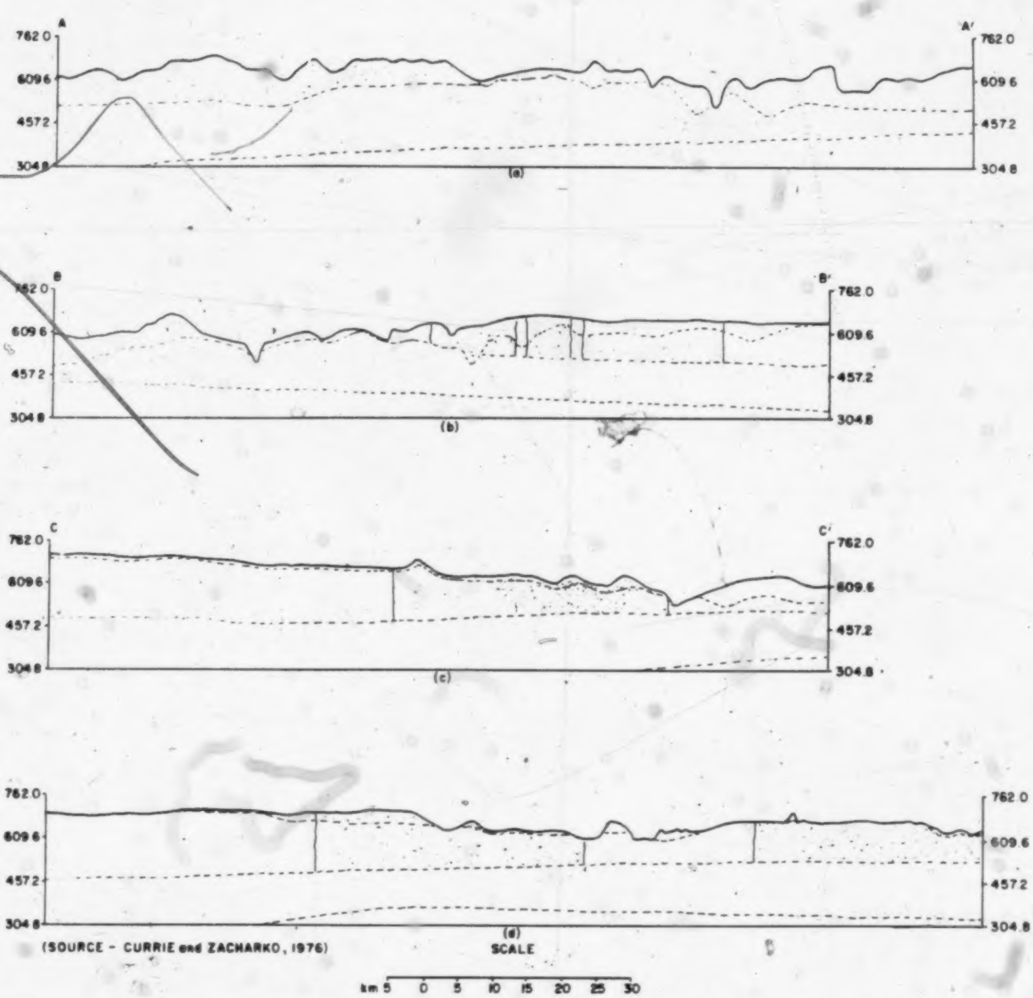
A'
metres
792.5
670.6
548.6
426.7
(a)
B
731.5
548.6
426.7
B'
731.5
548.6
426.7
(b)
C
792.5
670.6
548.6
426.7
C'
792.5
670.6
548.6
426.7
(c)
D
792.5
670.6
548.6
426.7
D'
792.5
670.6
548.6
426.7
(d)
SCALE
km 5
0
5
10
15 km

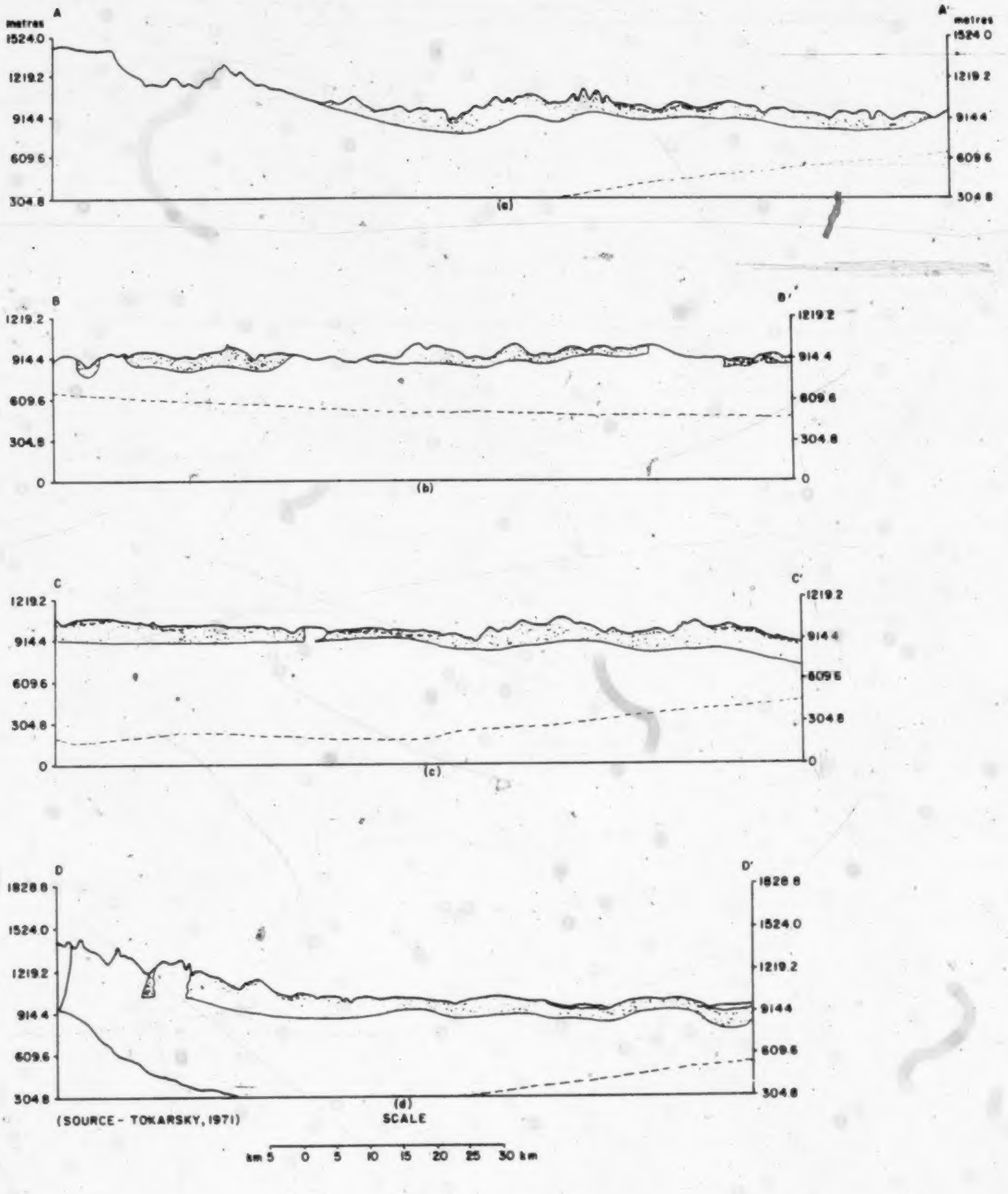
(SOURCE – STEIN, 1979)

Plate 5


A
A'
762.0
609.6
457.2
304.8
(a)
B
B'
(b)
C
C'
(c)
(d)
SCALE
km 5 0 5 10 15 20 25 30

(SOURCE - CURRIE and ZACHARKO, 1976)

Plate 6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1524.0
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1219.2
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304.8
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B'
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914.4
609.6
304.8
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(b)
C
1219.2
914.4
609.6
304.8
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D
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1524.0
1219.2
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1219.2
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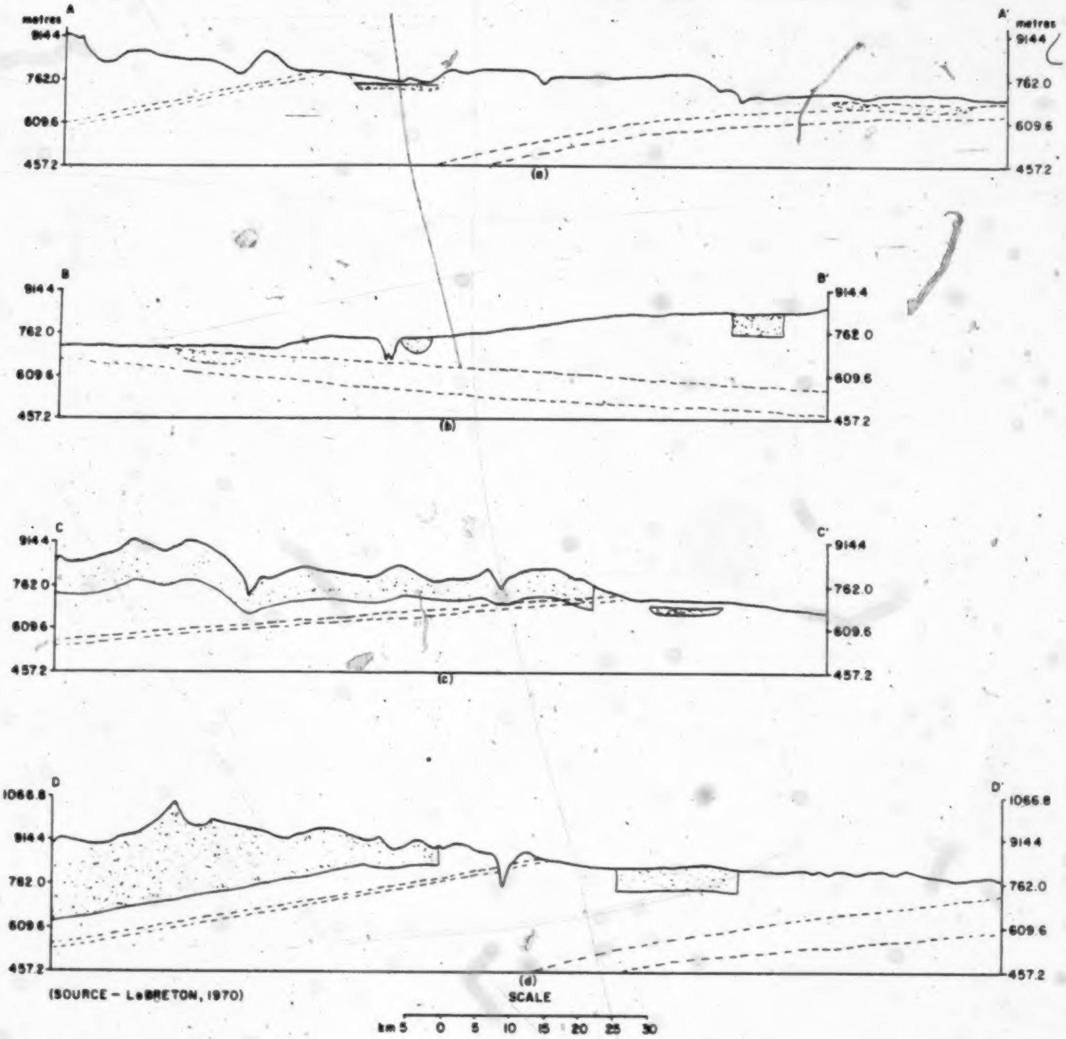
609.6
304.8
(d)
SCALE
km 5 0 5 10 15 20 25 30 km

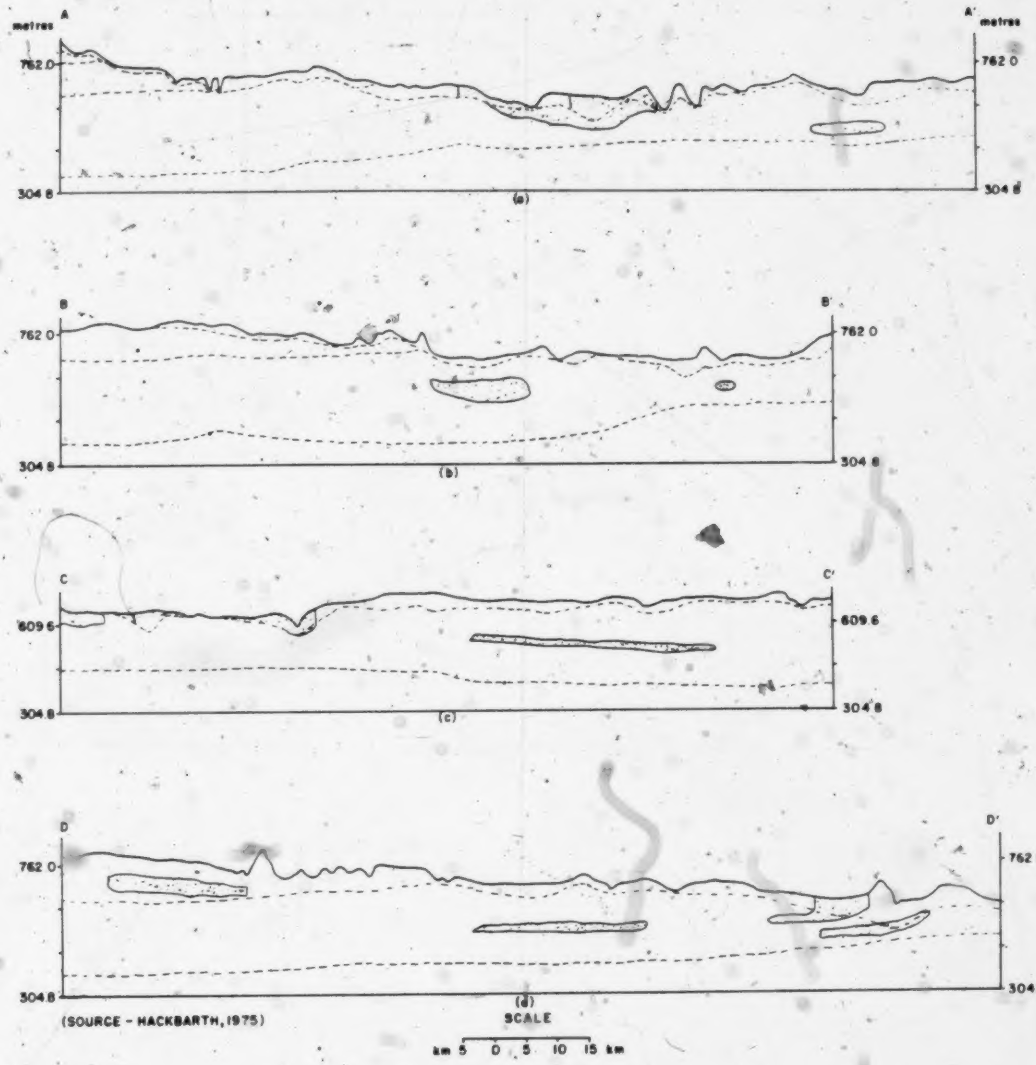
(SOURCE - TOKARSKY, 1971)

Plate 7


A
metres
914.4
762.0
609.6
457.2
(a)
A'
metres
914.4
762.0
609.6
457.2
B
914.4
762.0
609.6
457.2
(b)
B'
914.4
762.0
609.6
457.2
C
914.4
762.0
609.6
457.2
(c)
C'
914.4
762.0
609.6
457.2
D
1066.8
914.4
762.0
609.6
457.2
(d)
D'
1066.8
914.4
762.0
609.6
457.2
SCALE
km 5 0 5 10 15 20 25 30

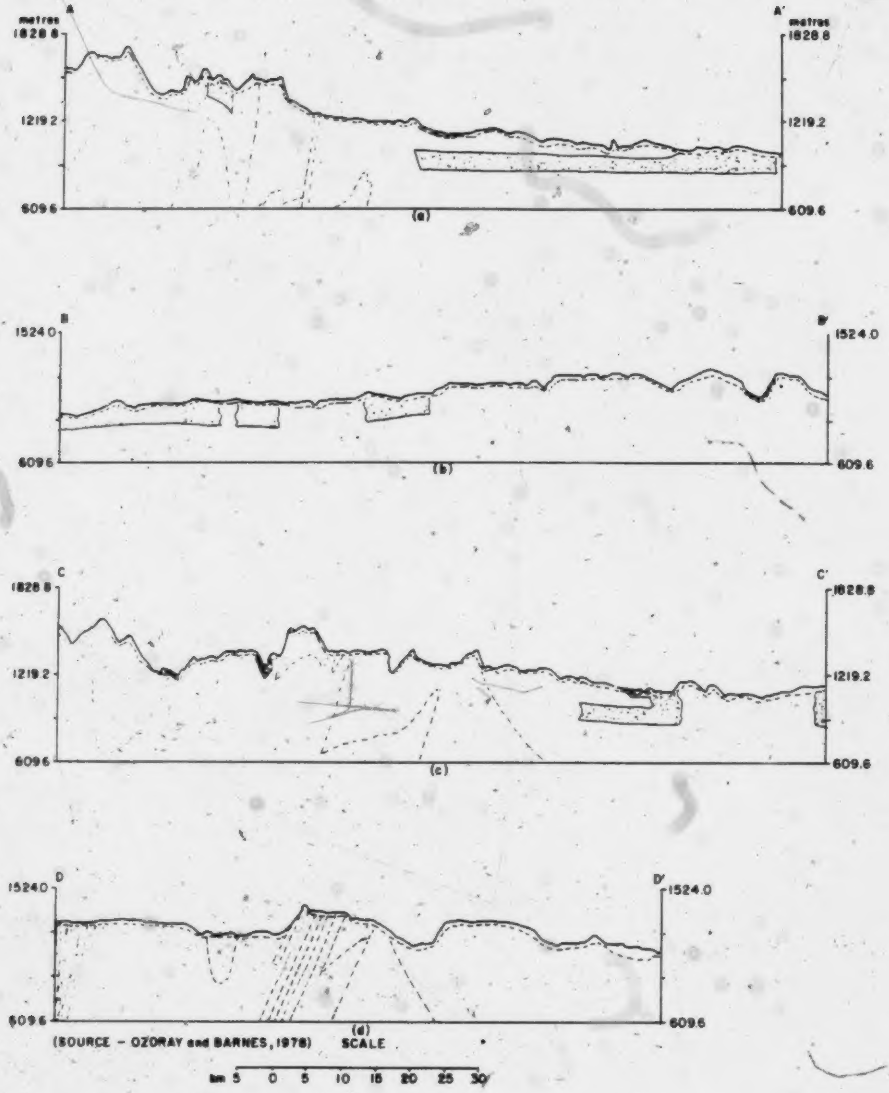
(SOURCE – LeBRETON, 1970)

Plate 8


A
metres
A'
metres
762.0
762.0
304.8
304.8
(a)
B
B'
762.0
762.0
304.8
304.8
(b)
C
C'
609.6
609.6
304.8
304.8
(c)
D
D'
762.0
762.0
304.8
304.8
(d)
SCALE
km 5 0 5 10 15 km

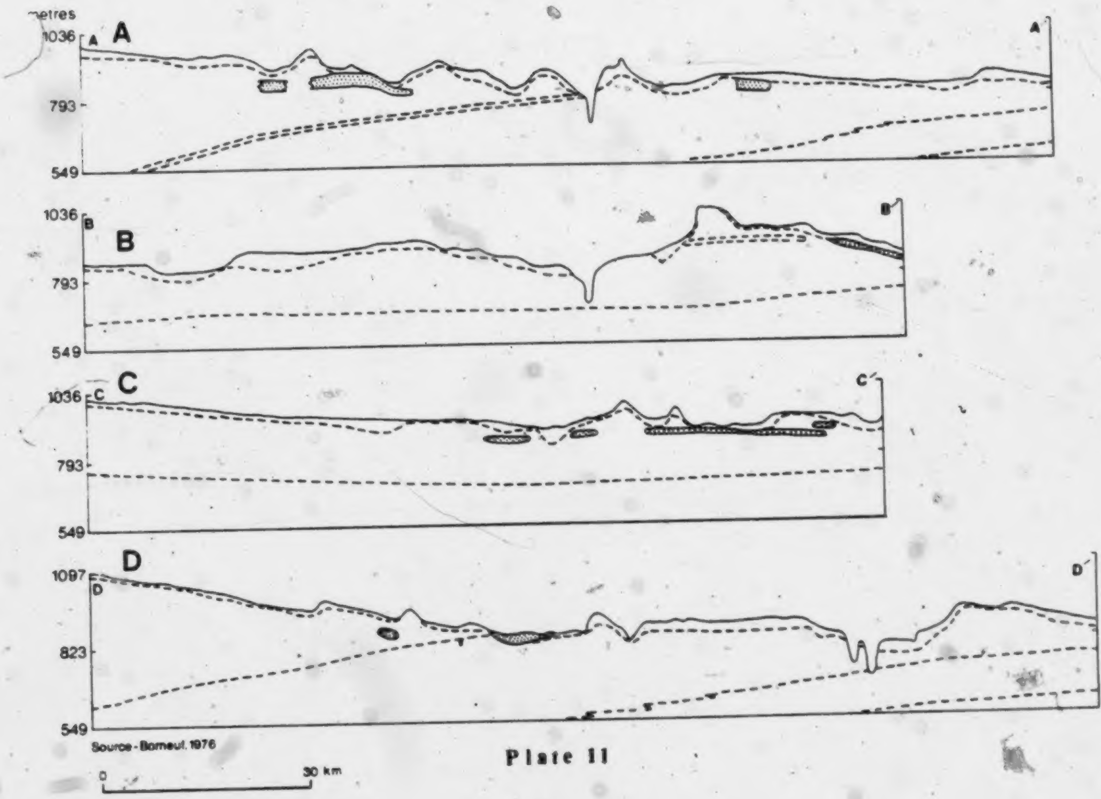
(SOURCE - HACKBARTH, 1975)

Plate 9


A
A'
metres
metres
1828.8
1828.8
1219.2
1219.2
609.6
609.6
(a)
B
B'
1524.0
1524.0
609.6
609.6
(b)
C
C'
1828.8
1828.8
1219.2
1219.2
609.6
609.6
(c)
D
D'
1524.0
1524.0
609.6
609.6
(d)
km 5 0 5 10 15 20 25 30

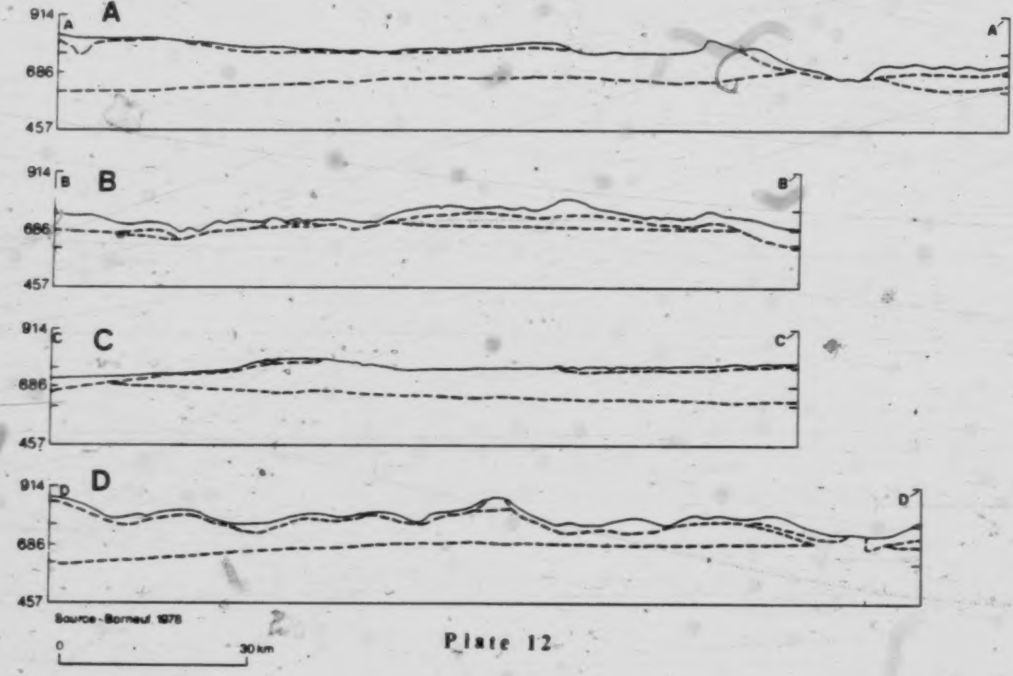
(SOURCE – OZORAY and BARNES, 1978) SCALE

Plate 10


metres
1036
793
549
A
A
A'
B
B
B'
1036
793
549
C
C
C'
1036
793
549
D
D
D'
1097
823
549
0
30 km

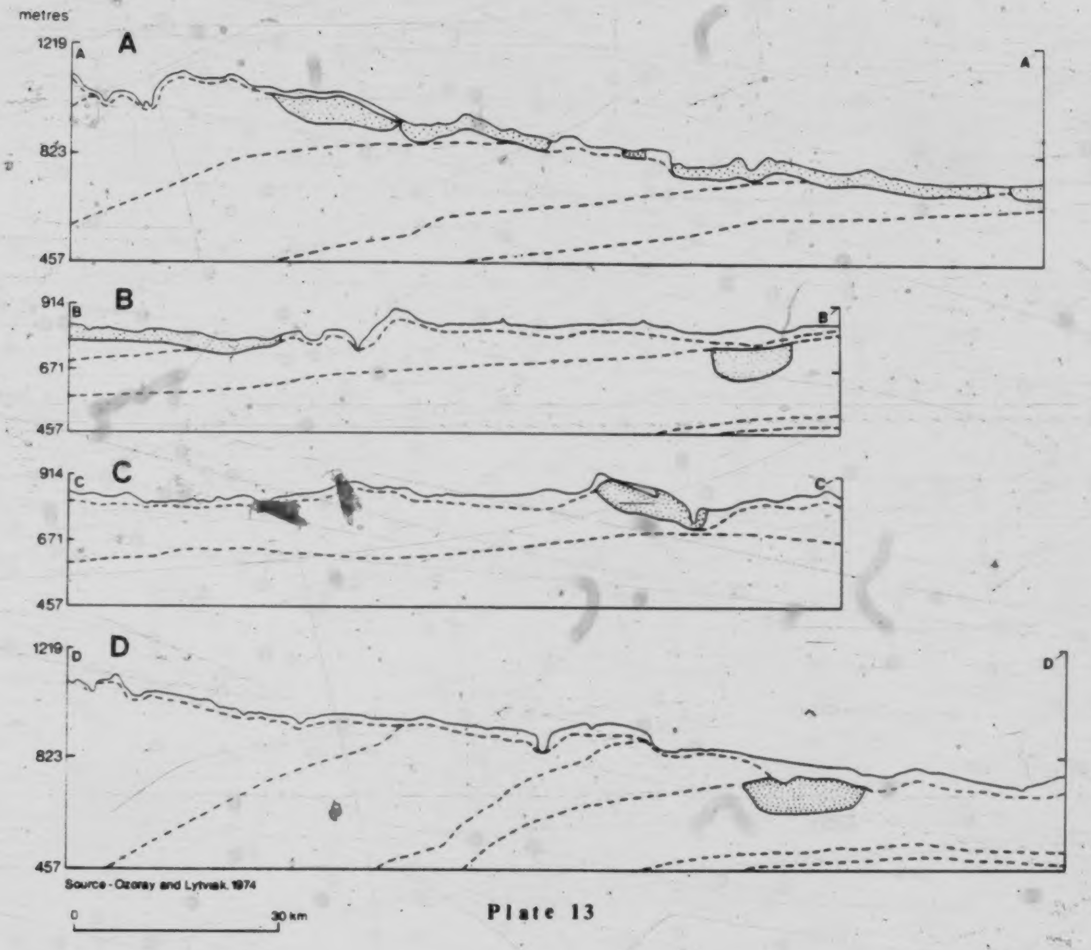
Source - Borneul, 1976

Plate 11



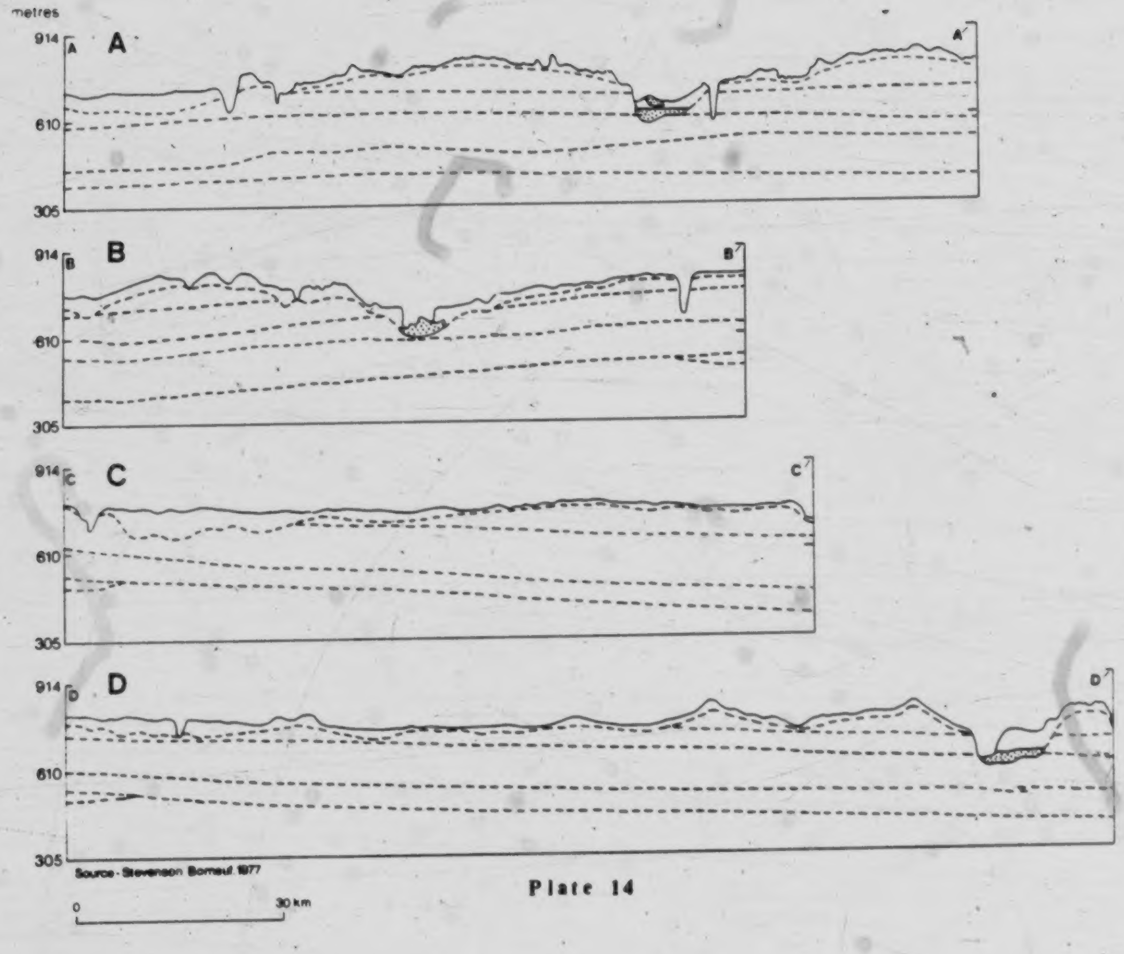
Source–Borneuf, 1978

Plate 12


metres
1219
823
457
A
A
A
914
671
457
B
B
B'
914
671
457
C
C
C'
1219
823
457
D
D
D'
0
30 km

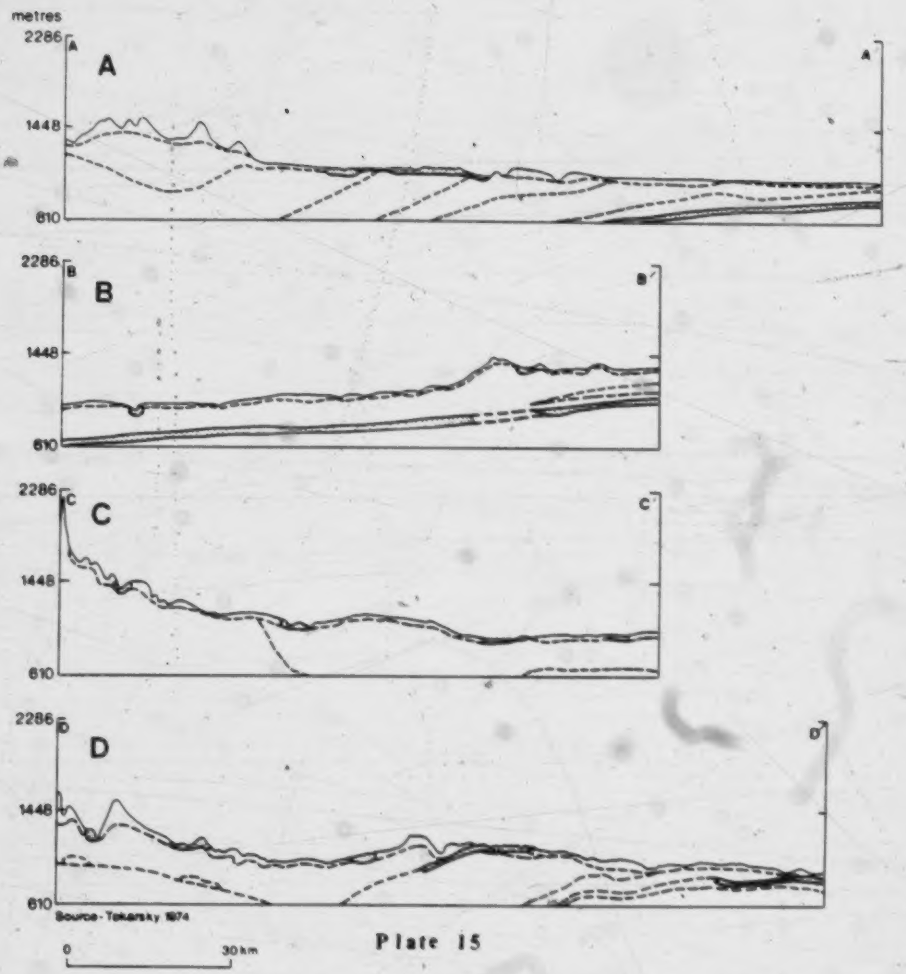
Source - Ozoray and Lytviak, 1974

**Plate 13**


metres
914
610
305
A
A'
B
B'
C
C'
D
D'
0
30 km

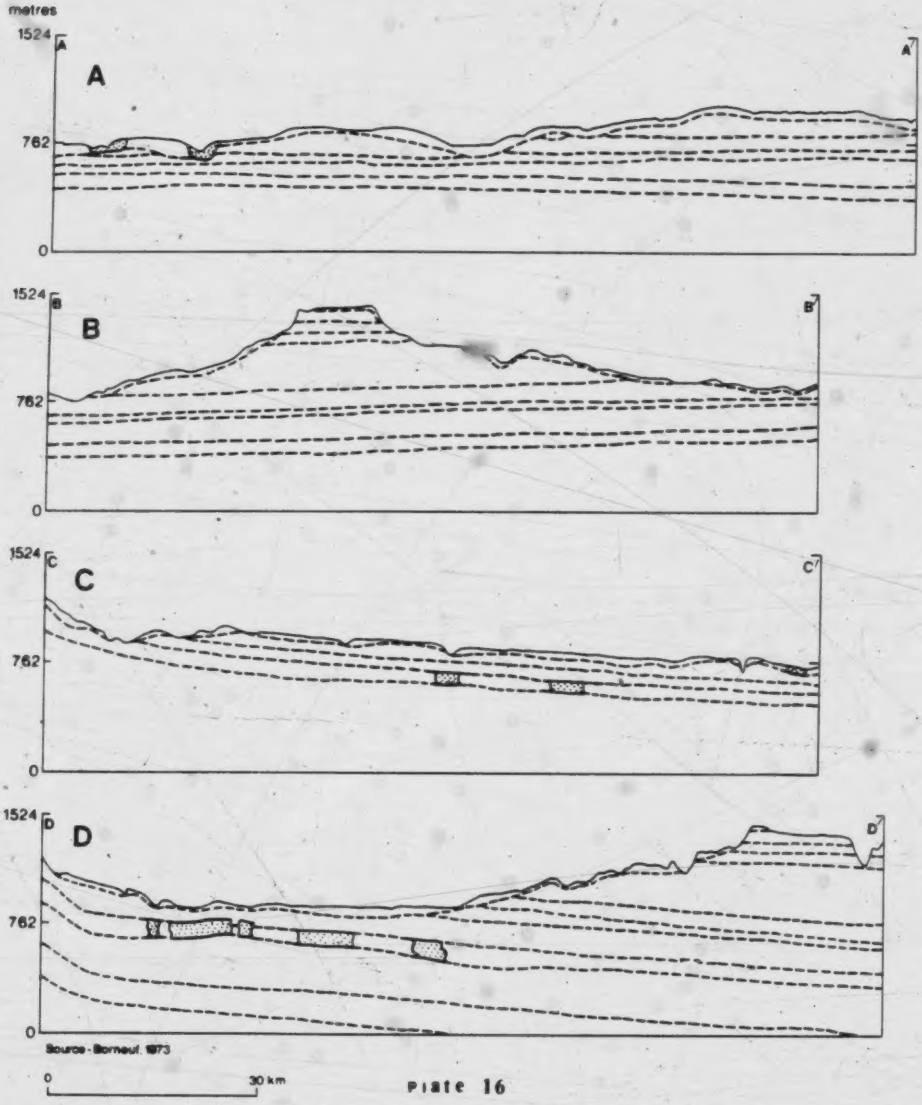
Source - Stevenson Borneuf 1977

Plate 14



Source - Tokarsky 1974

Plate 15


metres
1524
762
0
A
A'
B
B'
C
C'
D
D'
0
30 km

Source - Borneuf, 1973

Plate 16

## Appendix 2 - Simulated SBY and TBY values

### The main response curves

| Pumping distance from valley (x,m) | SBY (x 101.5 m/day) | TBY (x .519 $m^2$) | Remaining natural discharge (x 7.23 m/day) |
|---|---|---|---|
| 0.0 | 3.877 | 10.084 | 0.156 |
| 5.00000 | 2.397 | 9.862 | 1.326 |
| 10.00000 | 1.869 | 9.571 | 2.459 |
| 15.00000 | 2.057 | 9.218 | 3.545 |
| 20.00000 | 2.750 | 8.810 | 4.578 |
| 25.00000 | 3.764 | 8.353 | 5.547 |
| 30.00000 | 4.941 | 7.854 | 6.444 |
| 35.00000 | 6.145 | 7.318 | 7.260 |
| 40.00000 | 7.269 | 6.754 | 7.987 |
| 45.00000 | 8.229 | 6.166 | 8.615 |
| 50.00000 | 8.966 | 5.561 | 9.137 |
| 55.00000 | 9.453 | 4.947 | 9.544 |
| 60.00000 | 9.676 | 4.329 | 9.826 |
| 65.00000 | 9.656 | 3.714 | 9.975 |
| 70.00000 | 9.437 | 3.108 | 9.983 |
| 75.00000 | 9.087 | 2.518 | 9.840 |
| 80.00000 | 8.701 | 1.950 | 9.539 |
| 85.00000 | 8.398 | 1.412 | 9.069 |
| 90.00000 | 8.323 | 0.908 | 8.424 |

### Effect of well depth

| Pumping distance from valley (x, $10^3$m) | SBY (x 84.46 m/day) | TBY (x .79554 $m^2$) |
|---|---|---|
| | Deep wells | |
| 0.0 | 7.236 | 9.753 |
| 500.000 | 6.569 | 9.960 |
| 1000.000 | 6.331 | 9.904 |
| 1500.000 | 6.415 | 9.613 |
| 2000.000 | 6.727 | 9.117 |
| 2500.000 | 7.184 | 8.443 |
| 3000.000 | 7.715 | 7.619 |
| 3500.000 | 8.259 | 6.675 |
| 4000.000 | 8.767 | 5.637 |
| 4500.000 | 9.202 | 4.535 |
| 5000.000 | 9.539 | 3.397 |
| 5500.000 | 9.761 | 2.251 |
| 6000.000 | 9.867 | 1.126 |
| 6500.000 | 9.864 | 0.049 |
| 7000.000 | 9.772 | -0.950 |
| 7500.000 | 9.623 | -1.844 |
| 8000.000 | 9.458 | -2.604 |
| 8500.000 | 9.331 | -3.202 |
| 9000.000 | 9.309 | -3.610 |
| | Shallow wells | |
| 0.0 | 3.188 | 1.650 |
| 500.000 | 2.415 | 1.611 |
| 1000.000 | 2.141 | 1.571 |
| 1500.000 | 2.246 | 1.532 |
| 2000.000 | 2.622 | 1.492 |
| 2500.000 | 3.175 | 1.453 |
| 3000.000 | 3.822 | 1.413 |
| 3500.000 | 4.495 | 1.374 |
| 4000.000 | 5.138 | 1.334 |
| 4500.000 | 5.707 | 1.295 |
| 5000.000 | 6.173 | 1.256 |
| 5500.000 | 6.516 | 1.216 |
| 6000.000 | 6.734 | 1.177 |
| 6500.000 | 6.835 | 1.137 |
| 7000.000 | 6.838 | 1.098 |
| 7500.000 | 6.780 | 1.058 |
| 8000.000 | 6.706 | 1.019 |
| 8500.000 | 6.677 | 0.979 |
| 9000.000 | 6.766 | 0.940 |

### Effect of precipitation

| Pumping distance from valley (x, x $10^3$m) | SBY (x 26.72 m/day) | TBY (x .55049 $m^2$) |
|---|---|---|
| | Rate = .0242 m/day | |
| 0.0 | 8.223 | 10.000 |
| 500.000 | 7.797 | 9.900 |
| 1000.000 | 7.645 | 9.760 |
| 1500.000 | 7.700 | 9.590 |
| 2000.000 | 7.902 | 9.390 |
| 2500.000 | 8.196 | 9.170 |
| 3000.000 | 8.536 | 8.930 |
| 3500.000 | 8.886 | 8.670 |
| 4000.000 | 9.212 | 8.400 |
| 4500.000 | 9.491 | 8.120 |
| 5000.000 | 9.706 | 7.830 |
| 5500.000 | 9.849 | 7.530 |
| 6000.000 | 9.917 | 7.240 |
| 6500.000 | 9.916 | 6.950 |
| 7000.000 | 9.858 | 6.660 |
| 7500.000 | 9.764 | 6.380 |
| 8000.000 | 9.661 | 6.110 |
| 8500.000 | 9.584 | 5.860 |
| 9000.000 | 9.575 | 5.630 |
| | Rate = .018 m/day | |
| 0.0 | 7.310 | 9.500 |
| 500.000 | 6.795 | 9.440 |
| 1000.000 | 6.611 | 9.320 |
| 1500.000 | 6.674 | 9.140 |
| 2000.000 | 6.906 | 8.920 |
| 2500.000 | 7.241 | 8.660 |
| 3000.000 | 7.623 | 8.370 |
| 3500.000 | 8.004 | 8.040 |
| 4000.000 | 8.346 | 7.700 |
| 4500.000 | 8.622 | 7.330 |
| 5000.000 | 8.815 | 6.950 |
| 5500.000 | 8.915 | 6.560 |
| 6000.000 | 8.923 | 6.180 |
| 6500.000 | 8.851 | 5.800 |
| 7000.000 | 8.720 | 5.420 |
| 7500.000 | 8.559 | 5.070 |
| 8000.000 | 8.410 | 4.730 |
| 8500.000 | 8.320 | 4.420 |
| 9000.000 | 8.351 | 4.150 |
| | Rate = .0065 m/day | |
| 0.0 | 4.396 | 6.730 |
| 500.000 | 4.295 | 6.200 |
| 1000.000 | 4.285 | 5.670 |
| 1500.000 | 4.344 | 5.120 |
| 2000.000 | 4.448 | 4.560 |
| 2500.000 | 4.579 | 3.990 |
| 3000.000 | 4.721 | 3.420 |
| 3500.000 | 4.858 | 2.830 |
| 4000.000 | 4.979 | 2.230 |
| 4500.000 | 5.075 | 1.620 |
| 5000.000 | 5.140 | 1.000 |
| 5500.000 | 5.167 | 0.370 |
| 6000.000 | 5.157 | -0.270 |
| 6500.000 | 5.109 | -0.920 |
| 7000.000 | 5.027 | -1.580 |
| 7500.000 | 4.915 | -2.250 |
| 8000.000 | 4.784 | -2.930 |
| 8500.000 | 4.642 | -3.620 |
| 9000.000 | 4.502 | -4.320 |
| | Rate = .0017 m/day | |
| 0.0 | 1.066 | -13.440 |
| 500.000 | 1.068 | -13.510 |
| 1000.000 | 1.073 | -13.700 |
| 1500.000 | 1.080 | -13.990 |
| 2000.000 | 1.087 | -14.400 |
| 2500.000 | 1.092 | -14.920 |
| 3000.000 | 1.094 | -15.550 |
| 3500.000 | 1.093 | -16.280 |
| 4000.000 | 1.087 | -17.130 |
| 4500.000 | 1.077 | -18.090 |
| 5000.000 | 1.064 | -19.160 |
| 5500.000 | 1.048 | -20.340 |
| 6000.000 | 1.031 | -21.640 |
| 6500.000 | 1.013 | -23.040 |
| 7000.000 | 0.998 | -24.550 |
| 7500.000 | 0.986 | -26.170 |
| 8000.000 | 0.982 | -27.910 |
| 8500.000 | 0.988 | -29.750 |
| 9000.000 | 1.008 | -31.700 |

| Effect of slope | | |
|---|---|---|
| Pumping distance from valley (x, x $10^3$m) | SBY (x 51.74 m/day) | TBY (x .15553 m²) |
| | Slope = 0.05 | |
| 0.0 | 5.795 | 11.470 |
| 500.000 | 5.002 | 10.218 |
| 1000.000 | 4.842 | 8.966 |
| 1500.000 | 5.141 | 7.713 |
| 2000.000 | 5.749 | 6.461 |
| 2500.000 | 6.532 | 5.209 |
| 3000.000 | 7.379 | 3.957 |
| 3500.000 | 8.198 | 2.705 |
| 4000.000 | 8.916 | 1.452 |
| 4500.000 | 9.481 | 0.200 |
| 5000.000 | 9.861 | -1.052 |
| 5500.000 | 10.044 | -2.304 |
| 6000.000 | 10.037 | -3.556 |
| 6500.000 | 9.868 | -4.809 |
| 7000.000 | 9.584 | -6.061 |
| 7500.000 | 9.253 | -7.313 |
| 8000.000 | 8.963 | -8.565 |
| 8500.000 | 8.820 | -9.817 |
| 9000.000 | 8.953 | -11.070 |
| | Slope = 0.03 | |
| 0.0 | 4.067 | 9.769 |
| 500.000 | 3.406 | 9.976 |
| 1000.000 | 3.170 | 9.920 |
| 1500.000 | 3.253 | 9.629 |
| 2000.000 | 3.563 | 9.133 |
| 2500.000 | 4.016 | 8.459 |
| 3000.000 | 4.542 | 7.635 |
| 3500.000 | 5.081 | 6.691 |
| 4000.000 | 5.585 | 5.653 |
| 4500.000 | 6.017 | 4.551 |
| 5000.000 | 6.350 | 3.413 |
| 5500.000 | 6.571 | 2.267 |
| 6000.000 | 6.677 | 1.142 |
| 6500.000 | 6.675 | 0.065 |
| 7000.000 | 6.585 | -0.934 |
| 7500.000 | 6.439 | -1.828 |
| 8000.000 | 6.278 | -2.588 |
| 8500.000 | 6.155 | -3.186 |
| 9000.000 | 6.137 | -3.594 |
| | Slope = 0.0016 | |
| 0.0 | 1.241 | 2.570 |
| 500.000 | 1.150 | 2.537 |
| 1000.000 | 1.119 | 2.503 |
| 1500.000 | 1.133 | 2.470 |
| 2000.000 | 1.178 | 2.437 |
| 2500.000 | 1.243 | 2.403 |
| 3000.000 | 1.316 | 2.370 |
| 3500.000 | 1.389 | 2.337 |
| 4000.000 | 1.456 | 2.303 |
| 4500.000 | 1.512 | 2.270 |
| 5000.000 | 1.552 | 2.237 |
| 5500.000 | 1.577 | 2.203 |
| 6000.000 | 1.586 | 2.170 |
| 6500.000 | 1.580 | 2.137 |
| 7000.000 | 1.565 | 2.103 |
| 7500.000 | 1.546 | 2.070 |
| 8000.000 | 1.530 | 2.037 |
| 8500.000 | 1.526 | 2.003 |
| 9000.000 | 1.546 | 1.970 |

| Effect of basin depth | | |
|---|---|---|
| Pumping distance from valley (x, x $10^3$m) | SBY (x 10.02 m/day) | TBY (x .26075 m²) |
| | Depth/width = 1/10 (Deep basin) | |
| 0.0 | 8.086 | 9.769 |
| 500.000 | 7.626 | 9.976 |
| 1000.000 | 7.461 | 9.920 |
| 1500.000 | 7.520 | 9.629 |
| 2000.000 | 7.736 | 9.133 |
| 2500.000 | 8.052 | 8.459 |
| 3000.000 | 8.420 | 7.635 |
| 3500.000 | 8.796 | 6.691 |
| 4000.000 | 9.148 | 5.653 |
| 4500.000 | 9.450 | 4.551 |
| 5000.000 | 9.683 | 3.413 |
| 5500.000 | 9.838 | 2.267 |
| 6000.000 | 9.913 | 1.142 |
| 6500.000 | 9.912 | 0.065 |
| 7000.000 | 9.851 | -0.934 |
| 7500.000 | 9.750 | -1.828 |
| 8000.000 | 9.639 | -2.588 |
| 8500.000 | 9.556 | -3.186 |
| 9000.000 | 9.545 | -3.594 |
| | Depth/width = 1/50 (Shallow basin) | |
| 0.0 | 2.078 | 4.760 |
| 500.000 | 1.888 | 4.676 |
| 1000.000 | 1.778 | 4.592 |
| 1500.000 | 1.739 | 4.508 |
| 2000.000 | 1.763 | 4.424 |
| 2500.000 | 1.839 | 4.341 |
| 3000.000 | 1.958 | 4.257 |
| 3500.000 | 2.112 | 4.173 |
| 4000.000 | 2.290 | 4.089 |
| 4500.000 | 2.483 | 4.005 |
| 5000.000 | 2.683 | 3.921 |
| 5500.000 | 2.878 | 3.837 |
| 6000.000 | 3.061 | 3.753 |
| 6500.000 | 3.222 | 3.669 |
| 7000.000 | 3.350 | 3.586 |
| 7500.000 | 3.438 | 3.502 |
| 8000.000 | 3.476 | 3.418 |
| 8500.000 | 3.454 | 3.334 |
| 9000.000 | 3.363 | 3.250 |

| Effect of water table depth | | |
|---|---|---|
| Pumping distance from valley (x, x $10^3$m) | SBY (x 14.01 m/day) | TBY (x .33748 m³) |
| | Shallow water table | |
| 0.0 | 2.793 | 9.769 |
| 500.000 | 1.872 | 9.976 |
| 1000.000 | 1.543 | 9.920 |
| 1500.000 | 1.661 | 9.629 |
| 2000.000 | 2.093 | 9.133 |
| 2500.000 | 2.726 | 8.459 |
| 3000.000 | 3.459 | 7.635 |
| 3500.000 | 4.210 | 6.691 |
| 4000.000 | 4.910 | 5.653 |
| 4500.000 | 5.509 | 4.551 |
| 5000.000 | 5.968 | 3.413 |
| 5500.000 | 6.269 | 2.267 |
| 6000.000 | 6.406 | 1.142 |
| 6500.000 | 6.390 | 0.065 |
| 7000.000 | 6.248 | -0.934 |
| 7500.000 | 6.022 | -1.828 |
| 8000.000 | 5.771 | -2.588 |
| 8500.000 | 5.569 | -3.186 |
| 9000.000 | 5.506 | -3.594 |
| | Deep water tab[illegible] | |
| 0.0 | 7.947 | 43.730 |
| 500.000 | 5.886 | 42.904 |
| 1000.000 | 4.871 | 42.079 |
| 1500.000 | 4.647 | 41.253 |
| 2000.000 | 4.987 | 40.428 |
| 2500.000 | 5.693 | 39.602 |
| 3000.000 | 6.595 | 38.777 |
| 3500.000 | 7.550 | 37.951 |
| 4000.000 | 8.445 | 37.126 |
| 4500.000 | 9.193 | 36.300 |
| 5000.000 | 9.737 | 35.474 |
| 5500.000 | 10.048 | 34.649 |
| 6000.000 | 10.125 | 33.823 |
| 6500.000 | 9.994 | 32.998 |
| 7000.000 | 9.711 | 32.172 |
| 7500.000 | 9.360 | 31.347 |
| 8000.000 | 9.051 | 30.521 |
| 8500.000 | 8.925 | 29.696 |
| 9000.000 | 9.151 | 28.870 |

| Effect of hydraulic conductivity | | |
|---|---|---|
| Pumping distance from valley (x, x $10^3$m) | SBY (x 30.6 m/day) | TBY (x .47022 m³) |
| | K = 24.2 m/day | |
| 0.0 | 10.000 | -50.250 |
| 500.000 | 10.000 | -53.844 |
| 1000.000 | 10.000 | -57.229 |
| 1500.000 | 10.000 | -60.403 |
| 2000.000 | 10.000 | -63.367 |
| 2500.000 | 10.000 | -66.120 |
| 3000.000 | 10.000 | -68.664 |
| 3500.000 | 10.000 | -70.997 |
| 4000.000 | 10.000 | -73.120 |
| 4500.000 | 10.000 | -75.033 |
| 5000.000 | 10.000 | -76.736 |
| 5500.000 | 10.000 | -78.228 |
| 6000.000 | 10.000 | -79.511 |
| 6500.000 | 10.000 | -80.583 |
| 7000.000 | 10.000 | -81.445 |
| 7500.000 | 10.000 | -82.096 |
| 8000.000 | 10.000 | -82.538 |
| 8500.000 | 10.000 | -82.769 |
| 9000.000 | 10.000 | -82.790 |
| | K = .242 m/day | |
| 0.0 | 3.927 | 7.956 |
| 500.000 | 3.752 | 7.868 |
| 1000.000 | 3.690 | 7.703 |
| 1500.000 | 3.711 | 7.470 |
| 2000.000 | 3.791 | 7.176 |
| 2500.000 | 3.905 | 6.827 |
| 3000.000 | 4.036 | 6.432 |
| 3500.000 | 4.166 | 5.998 |
| 4000.000 | 4.284 | 5.532 |
| 4500.000 | 4.379 | 5.041 |
| 5000.000 | 4.446 | 4.533 |
| 5500.000 | 4.481 | 4.016 |
| 6000.000 | 4.486 | 3.497 |
| 6500.000 | 4.462 | 2.982 |
| 7000.000 | 4.418 | 2.481 |
| 7500.000 | 4.364 | 1.999 |
| 8000.000 | 4.313 | 1.545 |
| 8500.000 | 4.282 | 1.126 |
| 9000.000 | 4.290 | 0.748 |
| | K = .00242 m/day | |
| 0.0 | 1.002 | 9.927 |
| 500.000 | 1.008 | 9.848 |
| 1000.000 | 1.015 | 9.747 |
| 1500.000 | 1.022 | 9.627 |
| 2000.000 | 1.029 | 9.490 |
| 2500.000 | 1.037 | 9.338 |
| 3000.000 | 1.045 | 9.173 |
| 3500.000 | 1.053 | 8.998 |
| 4000.000 | 1.061 | 8.814 |
| 4500.000 | 1.070 | 8.624 |
| 5000.000 | 1.078 | 8.429 |
| 5500.000 | 1.086 | 8.232 |
| 6000.000 | 1.094 | 8.035 |
| 6500.000 | 1.102 | 7.840 |
| 7000.000 | 1.109 | 7.649 |
| 7500.000 | 1.116 | 7.464 |
| 8000.000 | 1.123 | 7.287 |
| 8500.000 | 1.129 | 7.121 |
| 9000.000 | 1.134 | 6.967 |

| Effect of extensive aquifers | | |
|---|---|---|
| Pumping distance from valley (x, x $10^3$m) | SBY (x 85.1 m/day) | TBY (x .80422 $m^3$) |
| | Deep aquifer | |
| 0.0 | 1.000 | -6.850 |
| 1000.000 | 3.340 | -8.420 |
| 2000.000 | 4.800 | -9.890 |
| 3000.000 | 5.700 | -11.270 |
| 4000.000 | 6.270 | -12.520 |
| 5000.000 | 6.600 | -13.610 |
| 6000.000 | 6.920 | -14.450 |
| 7000.000 | 6.640 | -14.890 |
| 8000.000 | 5.580 | -14.950 |
| 9000.000 | 3.730 | -15.000 |
| | Shallow aquifer | |
| 0.0 | 3.220 | 10.000 |
| 1000.000 | 2.600 | 9.330 |
| 2000.000 | 2.740 | 8.360 |
| 3000.000 | 4.640 | 7.180 |
| 4000.000 | 6.290 | 5.770 |
| 5000.000 | 7.900 | 4.400 |
| 6000.000 | 9.350 | 3.110 |
| 7000.000 | 9.950 | 1.980 |
| 8000.000 | 10.000 | 1.230 |
| 9000.000 | 9.700 | 1.000 |

| Effect of restricted aquifers | | |
|---|---|---|
| Pumping distance from valley (x, x $10^3$m) | SBY (x 46.6 m/day) | TBY (x .8986 $m^3$) |
| | Extensive aquifer | |
| 0.0 | 8.720 | 5.780 |
| 1000.000 | 8.600 | 6.350 |
| 2000.000 | 8.630 | 5.730 |
| 3000.000 | 8.990 | 4.970 |
| 4000.000 | 9.300 | 4.070 |
| 5000.000 | 9.600 | 3.180 |
| 6000.000 | 9.880 | 2.360 |
| 7000.000 | 9.990 | 1.630 |
| 8000.000 | 10.000 | 1.150 |
| 9000.000 | 9.990 | 1.000 |
| | Aquifer in discharge area | |
| 0.0 | 1.000 | 9.990 |
| 1000.000 | 4.850 | 7.910 |
| 2000.000 | 6.170 | 7.000 |
| 3000.000 | 5.930 | 6.850 |
| 4000.000 | 4.490 | 7.480 |
| 5000.000 | 1.600 | 8.980 |
| 6000.000 | 1.730 | 8.590 |
| 7000.000 | 1.800 | 8.250 |
| 8000.000 | 1.830 | 7.950 |
| 9000.000 | 1.820 | 7.760 |
| | Aquifer in midline area | |
| 0.0 | 1.150 | 9.960 |
| 1000.000 | 1.180 | 9.800 |
| 2000.000 | 1.250 | 9.620 |
| 3000.000 | 4.860 | 6.600 |
| 4000.000 | 6.030 | 5.490 |
| 5000.000 | 5.870 | 5.330 |
| 6000.000 | 4.610 | 6.190 |
| 7000.000 | 1.550 | 8.440 |
| 8000.000 | 1.680 | 8.030 |
| 9000.000 | 1.700 | 7.800 |
| | Aquifer in recharge area | |
| 0.0 | 1.070 | 10.000 |
| 1000.000 | 1.170 | 9.830 |
| 2000.000 | 1.270 | 9.630 |
| 3000.000 | 1.290 | 9.400 |
| 4000.000 | 1.230 | 9.160 |
| 5000.000 | 4.870 | 5.180 |
| 6000.000 | 5.940 | 3.940 |
| 7000.000 | 5.810 | 3.770 |
| 8000.000 | 4.700 | 4.790 |
| 9000.000 | 1.450 | 8.000 |

Effect of permissible drawdown

| Pumping distance from valley (x, x 10³m) | TBY (x .1291 m³) |
|---|---|
| | Permissible drawdown = 1/3 |
| 0.0 | 9.753 |
| 500.000 | 9.960 |
| 1000.000 | 9.904 |
| 1500.000 | 9.613 |
| 2000.000 | 9.117 |
| 2500.000 | 8.443 |
| 3000.000 | 7.619 |
| 3500.000 | 6.675 |
| 4000.000 | 5.637 |
| 4500.000 | 4.535 |
| 5000.000 | 3.397 |
| 5500.000 | 2.251 |
| 6000.000 | 1.126 |
| 6500.000 | 0.049 |
| 7000.000 | -0.950 |
| 7500.000 | -1.844 |
| 8000.000 | -2.604 |
| 8500.000 | -3.202 |
| 9000.000 | -3.610 |
| | Permissible drawdown = 1/10 |
| 0.0 | 1.700 |
| 500.000 | 1.683 |
| 1000.000 | 1.665 |
| 1500.000 | 1.646 |
| 2000.000 | 1.626 |
| 2500.000 | 1.605 |
| 3000.000 | 1.582 |
| 3500.000 | 1.558 |
| 4000.000 | 1.533 |
| 4500.000 | 1.507 |
| 5000.000 | 1.480 |
| 5500.000 | 1.452 |
| 6000.000 | 1.422 |
| 6500.000 | 1.391 |
| 7000.000 | 1.359 |
| 7500.000 | 1.326 |
| 8000.000 | 1.292 |
| 8500.000 | 1.257 |
| 9000.000 | 1.220 |

Effect of precipitation

| Pumping distance from valley (x, x 10[illegible]m) | SBY (x 26.72 m/day) | TBY (x .55049 $m^3$) |
|---|---|---|
| | Rate = .0242 m/day | |
| 0.0 | 8.223 | 10.000 |
| 500.000 | 7.797 | 9.900 |
| 1000.000 | 7.645 | 9.760 |
| 1500.000 | 7.700 | 9.590 |
| 2000.000 | 7.902 | 9.390 |
| 2500.000 | 8.196 | 9.170 |
| 3000.000 | 8.536 | 8.930 |
| 3500.000 | 8.886 | 8.670 |
| 4000.000 | 9.212 | 8.400 |
| 4500.000 | 9.491 | 8.120 |
| 5000.000 | 9.706 | 7.830 |
| 5500.000 | 9.849 | 7.530 |
| 6000.000 | 9.917 | 7.240 |
| 6500.000 | 9.916 | 6.950 |
| 7000.000 | 9.858 | 6.660 |
| 7500.000 | 9.764 | 6.380 |
| 8000.000 | 9.661 | 6.110 |
| 8500.000 | 9.584 | 5.860 |
| 9000.000 | 9.575 | 5.630 |
| | Rate = .018 m/day | |
| 0.0 | 7.310 | 9.500 |
| 500.000 | 6.795 | 9.440 |
| 1000.000 | 6.611 | 9.320 |
| 1500.000 | 6.674 | 9.140 |
| 2000.000 | 6.906 | 8.920 |
| 2500.000 | 7.241 | 8.660 |
| 3000.000 | 7.623 | 8.370 |
| 3500.000 | 8.004 | 8.040 |
| 4000.000 | 8.346 | 7.700 |
| 4500.000 | 8.622 | 7.330 |
| 5000.000 | 8.815 | 6.950 |
| 5500.000 | 8.915 | 6.560 |
| 6000.000 | 8.923 | 6.180 |
| 6500.000 | 8.851 | 5.800 |
| 7000.000 | 8.720 | 5.420 |
| 7500.000 | 8.559 | 5.070 |
| 8000.000 | 8.410 | 4.730 |
| 8500.000 | 8.320 | 4.420 |
| 9000.000 | 8.351 | 4.150 |
| | Rate = .0065 m/day | |
| 0.0 | 4.396 | 6.730 |
| 500.000 | 4.295 | 6.200 |
| 1000.000 | 4.285 | 5.670 |
| 1500.000 | 4.344 | 5.120 |
| 2000.000 | 4.448 | 4.560 |
| 2500.000 | 4.579 | 3.990 |
| 3000.000 | 4.721 | 3.420 |
| 3500.000 | 4.858 | 2.830 |
| 4000.000 | 4.979 | 2.230 |
| 4500.000 | 5.075 | 1.620 |
| 5000.000 | 5.140 | 1.000 |
| 5500.000 | 5.167 | 0.370 |
| 6000.000 | 5.157 | -0.270 |
| 6500.000 | 5.109 | -0.920 |
| 7000.000 | 5.027 | -1.580 |
| 7500.000 | 4.915 | -2.250 |
| 8000.000 | 4.784 | -2.930 |
| 8500.000 | 4.642 | -3.620 |
| 9000.000 | 4.502 | -4.320 |
| | Rate = .0017 m/day | |
| 0.0 | 1.066 | -13.440 |
| 500.000 | 1.068 | -13.510 |
| 1000.000 | 1.073 | -13.700 |
| 1500.000 | 1.080 | -13.990 |
| 2000.000 | 1.087 | -14.400 |
| 2500.000 | 1.092 | -14.920 |
| 3000.000 | 1.094 | -15.550 |
| 3500.000 | 1.093 | -16.280 |
| 4000.000 | 1.087 | -17.130 |
| 4500.000 | 1.077 | -18.090 |
| 5000.000 | 1.064 | -19.160 |
| 5500.000 | 1.048 | -20.340 |
| 6000.000 | 1.031 | -21.640 |
| 6500.000 | 1.013 | -23.040 |
| 7000.000 | 0.998 | -24.550 |
| 7500.000 | 0.986 | -26.170 |
| 8000.000 | 0.982 | -27.910 |
| 8500.000 | 0.988 | -29.750 |
| 9000.000 | 1.008 | -31.700 |

# Appendix 3 - A study of groundwater resources development and management in Ross Creek Basin, Alberta

# A Study of Groundwater Resources Development and Management in Ross Creek Basin : Alberta

FINAL REPORT

Submitted to
Planning Division
Alberta Environment
by
Duke U. Ophori[1] and József Tóth[2]

April, 1986

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[1] Graduate Student

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# ABSTRACT

Field studies were conducted in order to determine the hydrogeologic environment and some aspects of the groundwater regime in Ross Creek Basin, Alberta. The regional hydraulic gradient is dominated by the north-south component, thus, the basin's groundwater flow can be analyzed in a two-dimensional vertical cross section. Aquifer hydraulics reveals two restricted aquifers that can produce a maximum of $4.1 \times 10^{-3}$ $m^3/s$ of groundwater in order to mine the groundwater in twenty years. A greater portion of the basin is an aquitard. Analysis of water level fluctuation in observation wells indicates that 113 mm or $3.0 \times 10^{-4}$ m/day of water moves through the basin. Hydrogeological mapping indicates that groundwater moves thorough local, intermediate and regional systems. Groundwater chemistry is controlled by groundwater movement to a large extent. Up to 50 percent of the groundwater is potable, and the remainder may be used for other purposes.

Based on the hydrogeologic environment, a numerical simulation analysis of the groundwater regime suggests that 118 mm or $3.0 \times 10^{-4}$ m/day of water flows through the basin. About 40 percent of the groundwater is recharged in the areas around the Cypress Hills, whereas the remaining 60 percent is recharged through sporadically distributed zones of restricted sizes. Whereas a large amount of water is recharged into the basin, the bulk hydraulic conductivity of the geologic materials is lower than one needed to conduct the water to possible sites of development. Thus, $4.4 \times 10^{-2}$ m/s of groundwater only, or 13 percent of the total demands ($3.3 \times 10^{-1}$ $m^3/s$) can be produced on sustained basis, if the water rights of downstream water users are not respected. If downstream users are considered, and they should, groundwater pumpage is insignificant to a centrally operated water system in the basin. Springs at intermediate and higher topographic elevations may be licensed for use by the local inhabitants.

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# 1. INTRODUCTION

## 1.1 Location of Ross Creek Basin

Ross Creek Basin is situated in southeastern Alberta, in Improvement District No. 1. It covers an area of approximately 608 $km^2$ southeast of the City of Medicine Hat, bounded on the north by the Ross Creek valley at the Trans-Canada Highway, and on the south by the Cypress Hills divide (Figure 1). The basin extends between longitudes 110° 11' and 110° 36' W, and between latitudes 49° 36' and 50° 2' N. In terms of the Dominion land-survey system used commonly in the central and western provinces of Canada, the area is located in townships 8, 9, 10, 11 and 12, ranges 2, 3, 4 and 5, west of the 4th meridian.

## 1.2 The problem

The northern two-third of Ross Creek Basin is in the driest portions of the province of Alberta. Strictly speaking, precipitation is low, evapotranspiration balance is little for the often dry years of this area (Ophori and Tóth, 1983).

According to the Koppen's classification system (Trewartha, 1954), Ross Creek Basin lies in the "BSk" climatic belt. This climatic type is defined as follows: "B": dry climate; excess of evaporation over precipitation; no surplus of water to maintain a high groundwater level, so that permanent streams cannot originate within the climate type. "S": semi arid or steppe type. "k": average annual temperature under 18° C. However, recent studies indicate the presence of water surplus in average years, and a water table that is sub-parallel to the surface topography (Ophori and Tóth, 1983), suggesting that the area may not lie strictly in the BSk climatic zone.

Despite the poor conditions of water supply, plants and animals still abound in this area, that require water for their diurnal subsistence. Water is required mainly for domestic, irrigation, and municipal purposes.

Concerned with the conditions of water in the basin, prior to 1980, the Water Resources Management Services of the Planning Division of Alberta Environment, Lethbridge identified the need to evaluate the following aspects of the basin's water resources: 1. field hydrology; 2. water use; 3. water supply; 4. water management. Consequently, in 1980, field hydrological activities were undertaken by the Lethbridge Technical Services Division, and the Planning Division with the cooperation of the Water Resources Management Branch. During

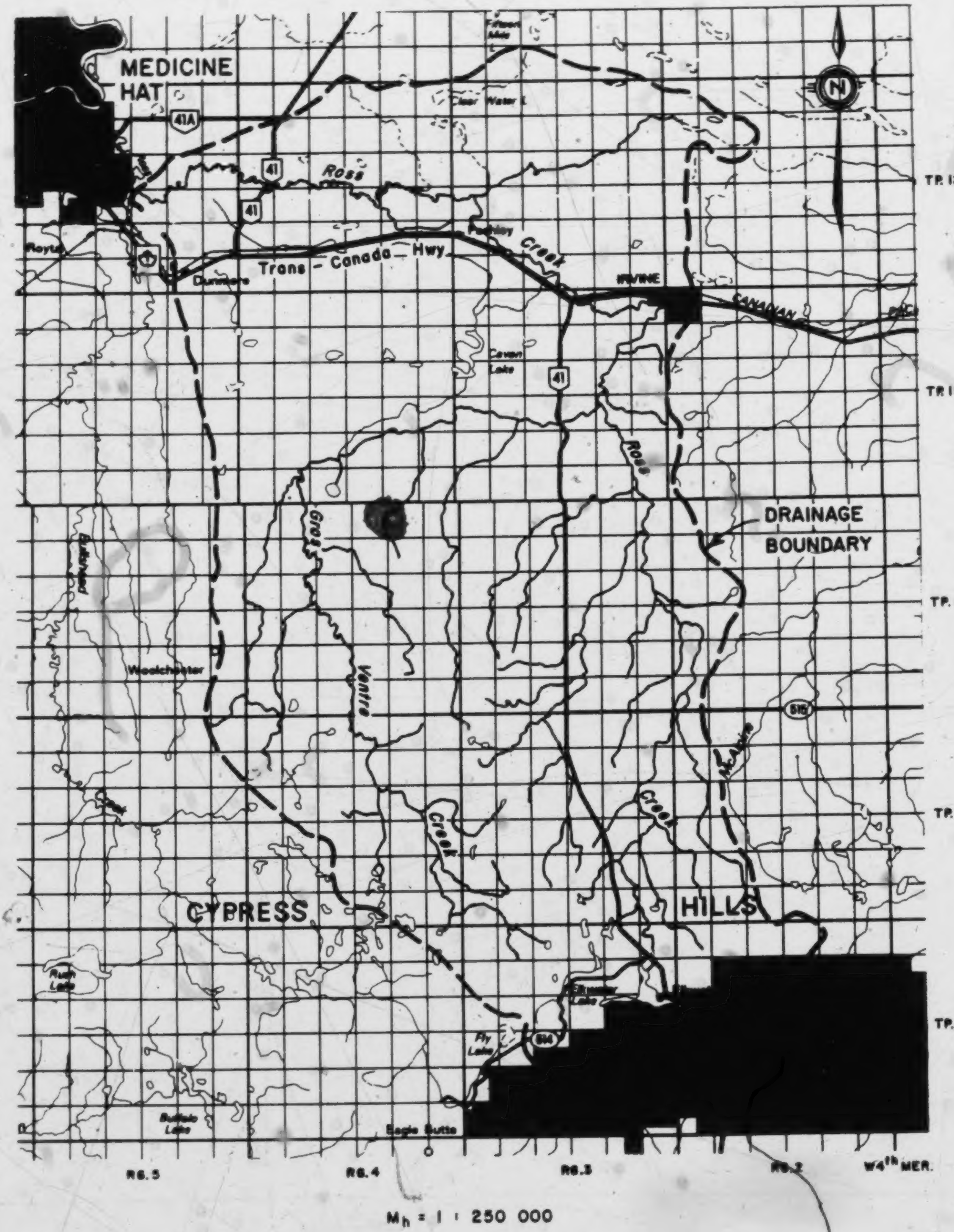

MEDICINE HAT
41A
41
Ross
Trans - Canada - Hwy
Dunmore
Royal
Fifteen Mile
Clear Water L
Pashley Creek
IRVINE
CANADIAN
Cavan Lake
TP. 12
TP. 11
TP. 10
TP. 9
TP. 8
Ross
DRAINAGE BOUNDARY
Gros
Ventre
Bullshead
Woolchester
Creek
Creek
Creek
McAlpine
515
CYPRESS
HILLS
Rush Lake
Elkwater Lake
Fly Lake
514
Buffalo Lake
Eagle Butte
RG.5
RG.4
RG.3
RG.2
W4th MER.
$M_h$ = 1 : 250 000

Figure 1. Location of Ross Creek Basin.

this study, drainage areas and existing data gathering networks were identified. To overcome existing data deficiencies, a five year data gathering program was proposed to provide data for analysis in the short and long term. Such data included snow surveys, meteorological station and soil moisture site installation, and the assessment of evaporation and channel losses in the area. Also, a preliminary water use survey was conducted along the main creek channels, and on irrigation. Following these developments, a hydrological assessment of the Ross Creek Basin was conducted by Northwest Hydraulic Consultants Ltd. (1981a, 1981b, 1983). These studies have reported the relatively high flows in Ross Creek and its major tributary, the Gros Ventre Creek in the spring months due to snow melt, and the drying up of these creeks along most of their courses during the remaining part of the year. In other words, the basin possesses a highly variable surface water supply.

It is clear from the preceding paragraph that most of the basin's water resources studies have been surface-water oriented. However, common knowledge shows that the water resources of a basin consist of both surface- and ground-water components. In spite of the dry nature of Ross Creek Basin, several springs discharge groundwater perennially at various topographic elevations and at various rates. Although the reliability of the springs is not known, their existence indicates a continuous supply of water to the groundwater system. Neither is the effect of spring development on the entire groundwater system known, thus, Alberta Environment is unwilling to issue licenses for their development as requested by the local inhabitants. The degree to which the entire groundwater system is interconnected determines its response to development at any particular site. Only when this degree of interconnection is known can a planned exploitation of the groundwater be made.

The problem at hand, therefore, is to evaluate the groundwater regime, that is: the occurrence, movement, quality, and possible use of the groundwater, and to provide a model for the management of the available groundwater resources in the basin.

### 1.3 Approach to the problem

The determination of the groundwater regime requires that the hydrogeologic environment of the basin be known. Thus, field studies were carried out with the major objective of ascertaining the hydrogeologic environment, and to obtain a few aspects of the groundwater regime. The resulting hydrogeologic environment was then used to evaluate the groundwater regime and its response to development theoretically.

## 2 LITERATURE AND FIELD STUDIES

### 2.1 Topography

The basin may be divided into three topographically distinct units: the Cypress Hills in the south, a dissected bench area in the central portion, and a relatively flat, low-lying land in the north, around the Trans-Canada Highway.

The flat-topped hills generally referred to as the Cypress Hills Plateaux by Westgate (1968), form a striking "linear high" along the southern margins of the basin and attain a maximum elevation of 1 463 m. At a very local scale, minor undulations are observed on the hill tops where local radial streams are fairly incised. The hills descend abruptly northwards into the dissected bench area at a rate of approximately 100 m per km. Consequently, a conspicuous break-of-slope shows up on aerial photographs of the junction between the hills and the bench area.

The "dissected bench" indicates the relatively high frequency of stream dissection which, together with their accompanying sloughs lend the upper bench area, a "knob-and-kettle" topographic outlook. The bench ranges from about 1 219 m to 915 m in elevation and slopes due north at an average rate of 14 m per km.

The flat land which occupies the northern part of the basin is gently undulating and has an average elevation of 793 m. The regional slope of this part of the basin is approximately 6 m per km, but the slope is interrupted by a prominent topographic feature: the U-shaped Ross Creek valley. Along this valley, the lowest point of the study area (below 693 m) lies to the northeastern corner, where Ross Creek flows out the basin toward the City of Medicine Hat.

Profiles of the east-west and north-south basin-wide topography are presented in Figure 2, the traces of which are indicated in Figure 1. From Figure 2, it may be concluded that Ross Creek Basin is uniform topographically along east-west profiles, whereas the south-north profile descends from viewing angle east to west, at the northern edge of the Cypress Hills Plateaux. The Cypress Hills area of the basin is an anomaly in the otherwise generally flat prairie landscape of Alberta.

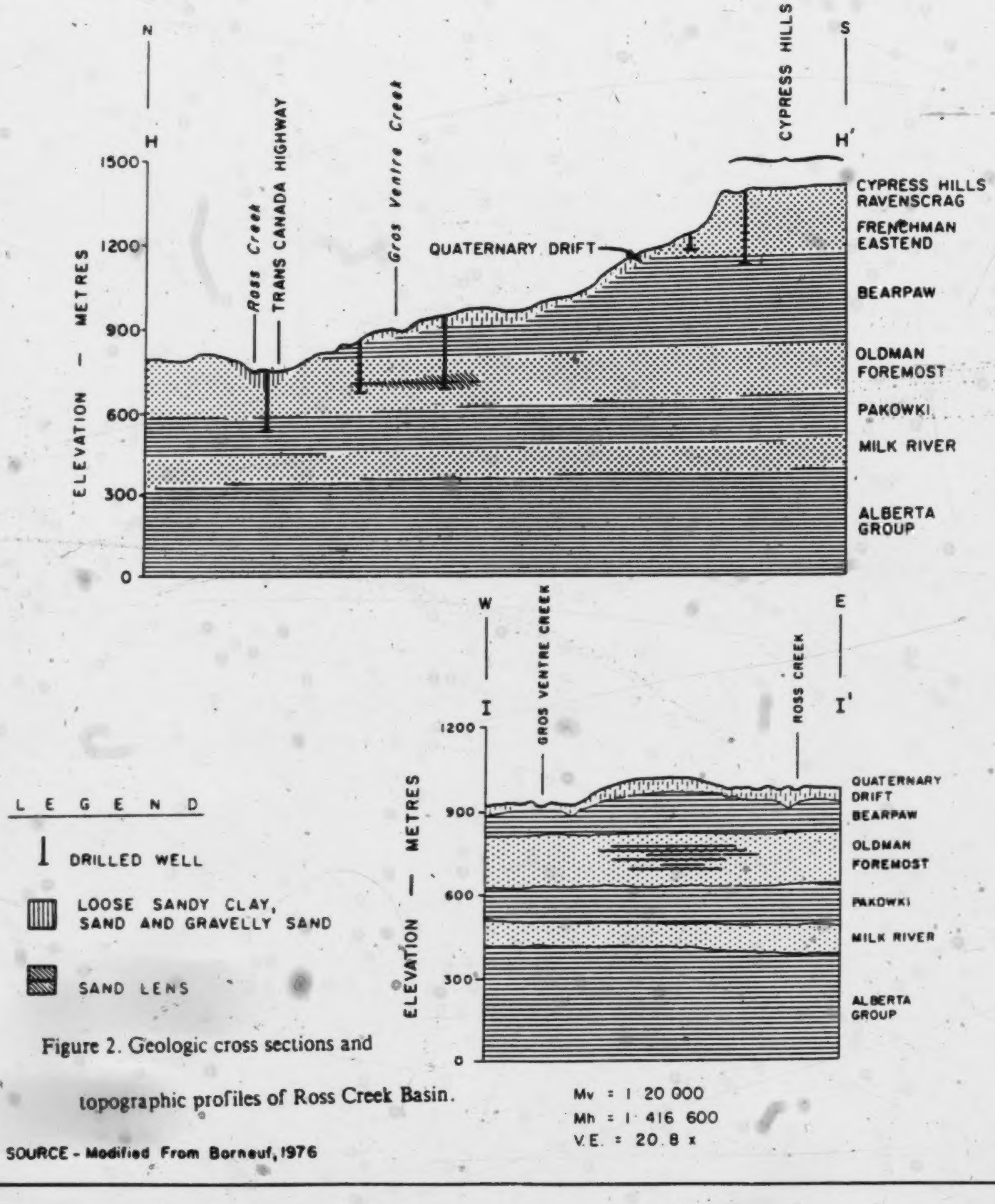

N
H
S
H'
CYPRESS HILLS
Ross Creek
TRANS CANADA HIGHWAY
Gros Ventre Creek
QUATERNARY DRIFT
ELEVATION — METRES
1500
1200
900
600
300
0
CYPRESS HILLS
RAVENSCRAG
FRENCHMAN
EASTEND
BEARPAW
OLDMAN
FOREMOST
PAKOWKI
MILK RIVER
ALBERTA GROUP
W
I
E
I'
GROS VENTRE CREEK
ROSS CREEK
QUATERNARY DRIFT
BEARPAW
OLDMAN
FOREMOST
PAKOWKI
MILK RIVER
ALBERTA GROUP
LEGEND
DRILLED WELL
LOOSE SANDY CLAY, SAND AND GRAVELLY SAND
SAND LENS
Mv = 1 20 000
Mh = 1 416 600
V.E. = 20.8 x

Figure 2. Geologic cross sections and topographic profiles of Ross Creek Basin.

SOURCE - Modified From Borneuf, 1976

### 2.2 Geology

The discovery of oil and gas in Alberta has led to large-scale subsurface geological work. Consequently, the subsurface geology of Alberta may be considered to be well defined on the regional scale. Subsurface exploration and historical geological studies have led to the recognition that the rocks and sediments found in Alberta record a history of events that began more than 2.6 billion years ago.

Rocks older than 590 million years belong to the Precambrian or Cryptozoic eon and give evidence of extensive periods of mountain building and volcanism. The younger sedimentary rocks stacked upon the crystalline Precambrian rocks record the history of periodic invasions and retreats of ancient seas, the uplift and erosion of land masses and mountains, and the marked changes in climate from tropic to arctic. Cycles of glacial erosion and deposition over the past 2 million years are marked by the surficial sediments which record the advances and ultimate stagnation and melting of continental and mountain glaciers.

The complex sequence of rocks and sediments is often divided into a stratigraphic section of four basic units: Precambrian bedrock, Paleozoic bedrock, Post-Paleozoic bedrock and surficial deposits. As groundwater studies are often limited to relatively shallow depths, only the Post-Paleozoic bedrock and surficial deposits are considered herein. Unfortunately, a common problem associated with groundwater studies in the area of study is that the oil industry is uninterested in these shallow rock units, hence detailed information about each rock unit is rare or, in most cases, nonexistent.

The geology as described is, mostly, a synthesis of previous larger scale studies by Russell and Landes (1940), Irish (1968), Green (1972), Borneuf (1976), and Stevenson and Borneuf (1977). Supplementary information was also obtained from the Groundwater Section of Alberta Environment, from the Coal Division of Energy Resources Conservation Board, and from a drilling program as a part of the present study. A composite stratigraphic section and generalized geologic cross sections are enclosed in Appendix 1.

#### 2.2.1 Bedrock

Ross Creek Basin is underlain by nearly flat-lying marine and non-marine layers of sandstones and shales of late Upper Cretaceous and Tertiary age (Figure 2).

The oldest known Cretaceous strata in the basin are those of the marine Alberta Group, consisting of dark grey, friable and fissile sandy shales and brown-weathering, grey sandstones. These strata are overlain by the Milk River Formation.

The lower member of the Milk River Formation consists of marine massive, light-grey to white weathering, grey, soft and hard sandstone. In addition, grey and light-grey shale, and sandy shale are also present in this member. Occurrence of ironstone concretions has been reported in the lower member which is the main aquifer in the western half of the Foremost map-area (Borneuf, 1976), beyond the limits of Ross Creek Basin. The upper member of the Milk River Formation is non-marine, and it consists essentially of softer argillaceous sandstones and sandy carbonaceous shales.

Overlying the Milk River strata are dark grey shales and sandy shales with some interbedded, thin, grey sandstone strata, comprising the marine Pakowki Formation. At places, the base of the formation is marked by a thin, chert-pebble conglomerate, or by a layer of small chert pebbles within the shale, indicating a disconformity at the Milk River-Pakowki contact.

The Pakowki shales grade upwards into the Foremost Formation. The latter comprises a succession of interbedded green and grey shales, dark grey carbonaceous shales, grey siltstones, and grey to pale brown sandstones. The carbonaceous shales grade, in some places, into coal seams, and ironbeds are common. The sandstones are mainly soft and argillaceous, but thin, hard layers and concretionary masses of calcareous buff weathering sandstone occur throughout. Individual units are thin lenses, and they do not extend far in any direction. In the middle part of the formation, *Ostrea* and *Corbula* form shell beds commonly.

The upper beds of the Foremost Formation are transitional into the overlying Oldman Formation so that, to separate the two formations, only an arbitrary contact is placed usually. The Oldman Formation crops out in most of the northern part of the basin (Figure 3), but it is overlain unconformably by Quaternary deposits, occasionally. Oldman strata comprise green, grey and light-grey shales, and silty shales interbedded and interlensed with grey and light grey sandstones. Throughout the succession, hard, calcareous sandstone beds and thin ironstone bands occur. The upper part of the formation contains thick, crossbedded, light-grey weathering, lenticular sandstone units. The formation is less carbonaceous, except that its top is marked by mineable coal seams throughout southern Alberta. Vertebrate remains and shells,

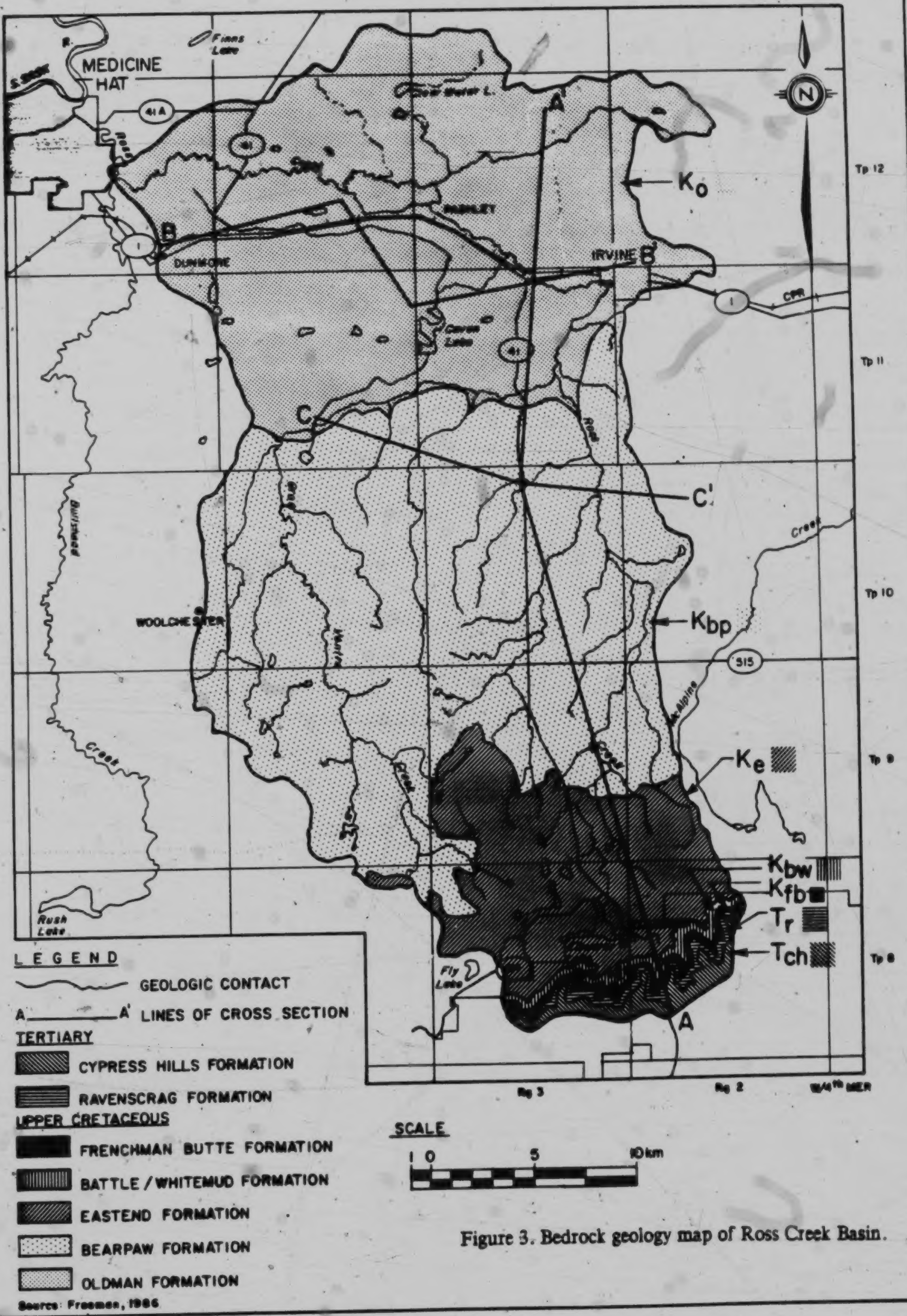

MEDICINE HAT
S. Sask. R.
Finns Lake
Sweet Water L.
41A
41
Ross
REDCLIFF
DUNMORE
IRVINE
B
B'
A
A'
C
C'
CPR
1
Cavan Lake
Ko
Kbp
Ke
Kbw
Kfb
Tr
Tch
N
Tp 12
Tp 11
Tp 10
Tp 9
Tp 8
Bullshead
Creek
WOOLCHESTER
Ross
Creek
McAlpine
315
Rush Lake
Fly Lake
Rg 3
Rg 2
W/4TH MER
LEGEND
GEOLOGIC CONTACT
A — A' LINES OF CROSS SECTION
TERTIARY
CYPRESS HILLS FORMATION
RAVENSCRAG FORMATION
UPPER CRETACEOUS
FRENCHMAN BUTTE FORMATION
BATTLE / WHITEMUD FORMATION
EASTEND FORMATION
BEARPAW FORMATION
OLDMAN FORMATION
Source: Freeman, 1986
SCALE
1 0 5 10 km

Figure 3. Bedrock geology map of Ross Creek Basin.

mainly Ostrea, occur in the formation. The Oldman Formation is an aquifer in a few locations along the northern portion of the Ross Creek Basin (Borneuf, 1976).

The marine Bearpaw Formation which rests conformably on the Oldman beds is the topmost bedrock unit in most parts of the basin but underlies the southern portion of the study area flanking the Cypress Hills (Figure 3). The formation consists of grey weathering, dark grey or brownish grey shale and silty shale with spheroidal-ironstone concretions, fine-grained clayey sandstone and benthonite beds. Thick sandstone units occur in the upper part of the formation.

Confined to the Cypress Hills area, the sandstone strata of the Eastend Formation overlie the Bearpaw shale unconformably. The formation consists of grey- to buff-weathering, grey and pale brown, fine- to medium-grained clayey sandstone; green, grey and dark grey shale; grey silty shale and siltstone; black carbonaceous shales and coal seams. A mineable coal seam occurs about 21 m below the top of the formation, in the vicinity of Elkwater Lake (Irish, 1968). Although the upper part is non-marine, marine fossils occur in some lower beds.

The Eastend Formation is overlain by the generally thin non-marine Whitemud Formation. The formation consists of white-weathering, fine- to medium-grained, felspathic, crossbedded argillaceous sandstones and white- to cream-weathering, grey, and pale green clays and silty clays, with a total thickness of about 8 m.

The dark clays and shales of the Battle Formation overlie the Whitemud Formation, conformably. The formation is very uniform and the shales are puplish, silty and benthonic with a grey-brown siliceous tuff bed occurring within the upper part. Its upper contact is an erosional surface resulting in a variable thickness up to a maximum of 9.1 m.

The Battle Formation is overlain by the Frenchman Formation consisting mainly of massive, medium-grained, crossbedded, buff- to reddish-brown, grey and light-brown sandstone intercalated with beds of green and grey siltstone, silty shale and shale.

The upper contact of the Frenchman Formation is transitional into the overlying Tertiary, non-marine Ravenscrag Formation, which is confined to a narrow rim around the higher parts of the Crypress Hills. The beds consist of soft, grey and light-brown, fine-grained sandstone and brownish grey clays. Lignitic coal seams, benthonic layers and thin beds of ironstone are often encountered within the formation.

The youngest bedrock unit within Ross Creek Basin is the Cypress Hills Formation which caps the Cypress Hills and consists mainly of a hard, well cemented, pebble and cobble conglomerate with some intercalated grey, coarse, sandstone lenses. A matrix of medium-grained, calcareous sandstone binds the brown and purplish chert and grey quartzite content of the pebbles and cobbles.

### 2.2.2 Surficial deposits

The surficial deposits encountered in Ross Creek Basin have been described variously in larger scale reports by different workers in southern Alberta and its environs. A general consensus of opinion about the surficial deposits is described in detail by Westgate (1965, 1968).

The general stratigraphic column of the Pleistocene deposits in the Foremost-Cypress Hills area is shown in Figure 4. According to Westgate (1968), only the Saskatchewan gravels, Wolf Island sediments, Cypress Hills loess, and Many Berries volcanic ash are rock-stratigraphic units as defined by the American Commission on Stratigraphic Nomenclature (1961, Art 4). The glacial drift sheets have been termed as morphostratigraphic units as defined by Frye and Willman (1960), because they have been differentiated on the basis of their topographic form, geographic position, and inferred geologic history. The occurrence of morphostratigraphic units in the Foremost-Cypress Hills area, is presented in Figure 5.

The Saskatchewan gravels are fluviatile sediments of early to late Pleistocene age. They occur as channel fill deposits in the preglacial valleys and range from sand to pebbles and cobbles of dominantly quartzite, argillite and chert. Their overall reddish-brown colour has resulted from stains by ferric oxides. Although the Saskatchewan gravels have not been reported at any locality within the Ross Creek Basin, they are known to lie upon upper Cretaceous sediments at the bottoms of major preglacial valleys in the Foremost-Cypress Hills area.

The Wolf Island sediments consist of lacustrine sands, and clays that overlie the Saskatchewan gravels conformably, within the major preglacial valleys. These sediments accumulated in cold, quite elongate preglacial lakes that formed along the preglacial valleys when the outlets of their rivers which then drained northeastwards into Hudson Bay, became blocked by the enchroaching Laurentide glacier.

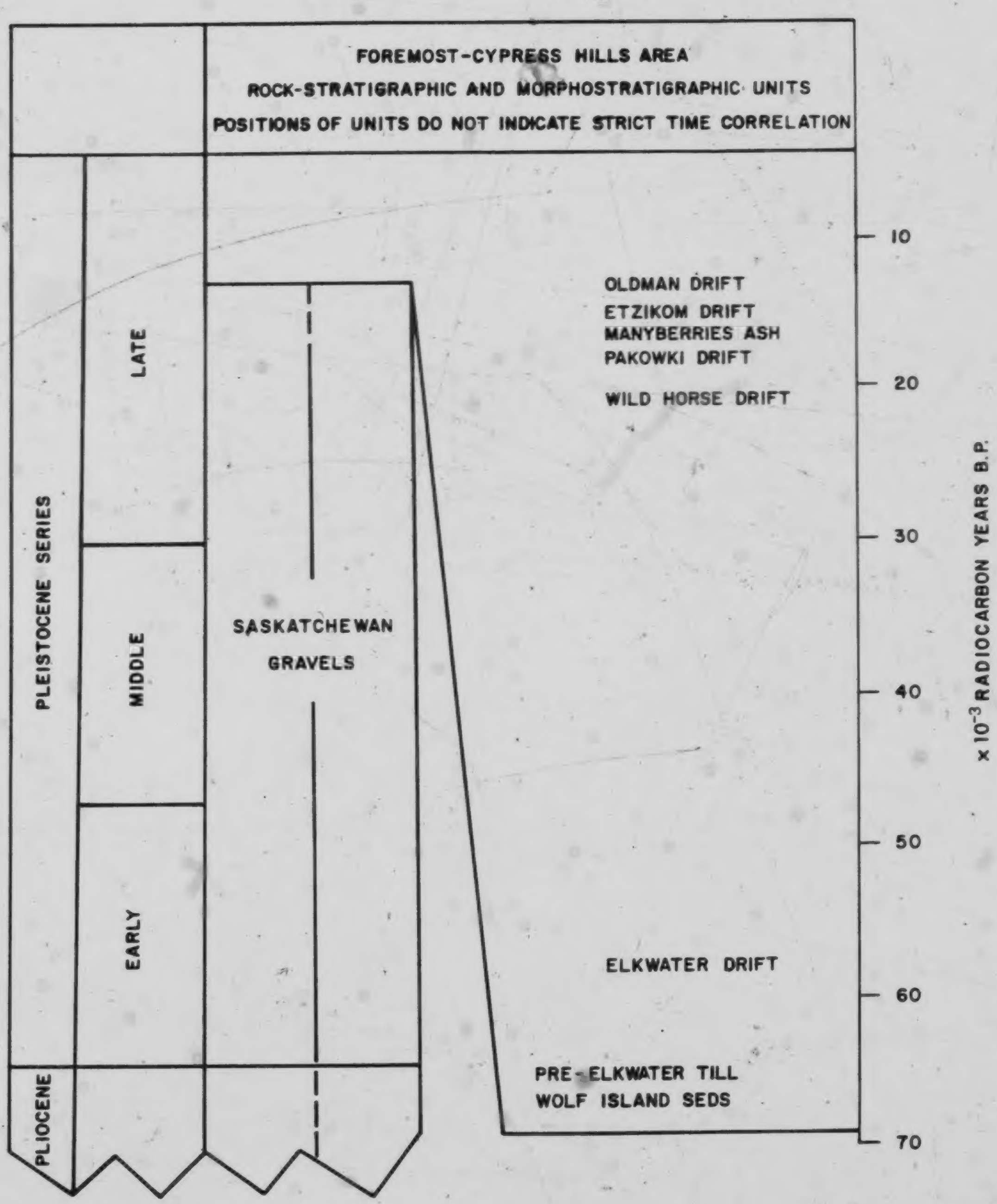

FOREMOST-CYPRESS HILLS AREA
ROCK-STRATIGRAPHIC AND MORPHOSTRATIGRAPHIC UNITS
POSITIONS OF UNITS DO NOT INDICATE STRICT TIME CORRELATION
PLEISTOCENE SERIES
PLIOCENE
LATE
MIDDLE
EARLY
SASKATCHEWAN GRAVELS
OLDMAN DRIFT
ETZIKOM DRIFT
MANYBERRIES ASH
PAKOWKI DRIFT
WILD HORSE DRIFT
ELKWATER DRIFT
PRE-ELKWATER TILL
WOLF ISLAND SEDS
10
20
30
40
50
60
70
x 10^-3 RADIOCARBON YEARS B.P.

Figure 4. Stratigraphic column of Quaternary deposits in the Foremost-Cypress Hills area.

SOURCE – Modified from Westgate, 1968

Four of the five surficial drift sheets (morphostratigraphic units) identified in the Foremost-Cypress Hills area, occur within Ross Creek Basin; Elkwater, Wild Horse, Pakowki, and Etzikom drifts. The Elkwater drift which is confined to the Cypress Hills and its confines (Figure 5) is thin, and it consists of sporadic patches of glacial gravel and till, together with some large erratics, including dolomite erratics. The gravels within the drift are composed of 87 percent quartzite pebbles, 9 percent granitic rocks from the Canadian Shield and 4 percent dolomite pebbles (Westgate, 1968). The Wild Horse drift consists predominantly of till, and contains extensive glacio-fluvial and glacio-lacustrine deposits. The till has a clay-loam to loam texture, and its pebble composition is 50 to 60 percent dolomite and limestone, with the remainder consisting dominantly of granitic rocks from the Canadian Shield. The Pawkowki drift consists mainly of till that has a loam texture, being somewhat coarser-grained than the Wild Horse drift. The topography is rugged, and undrained depressions are common as in the Wild Horse drift. The Etzikon drift covers most of the northern quarter of the basin and consists mainly of till with a high content of glacio-lacustrine sediments and crystalline rocks from the Canadian Shield. This drift, like the others, is believed to belong to the Wisconsin Stage.

The Cypress Hills loess occupies the Cypress Hills Plateaux, and overlies the weathered cap of the Cypress Hills conglomerate (Figure 5). The sandy loess has a massive structure and varies in thickness from 0.5 to 2.5 m. The most common minerals in the loess are quartz, mica, feldspar, opaques, garnet, epidote and fresh blue-green hornblende. The poorly sorted loess is coarse-grained in places and contains quartzite pebbles, a factor which would result in effective infiltration of rainwater and evaporation of soil moisture on top of the Cypress Hills.

### 2.2.3 Summary of geology

The relatively detailed description of the geology reveals that, in Ross Creek Basin, because of the periodic enchroachment of the sea on the land, the bedrock consists of alternating formations of supposedly high and low permeability. Some of the formations are regionally continuous throughout the basin. The high permeability formations consist mainly of sandstones intercalated with siltstones and shales, whereas the low permeability formations are dominantly shales with lenses of siltstone and sandstone. In general, the degree of interlayering of sandstone, siltstone, and shales for both high and low permeability formations approach

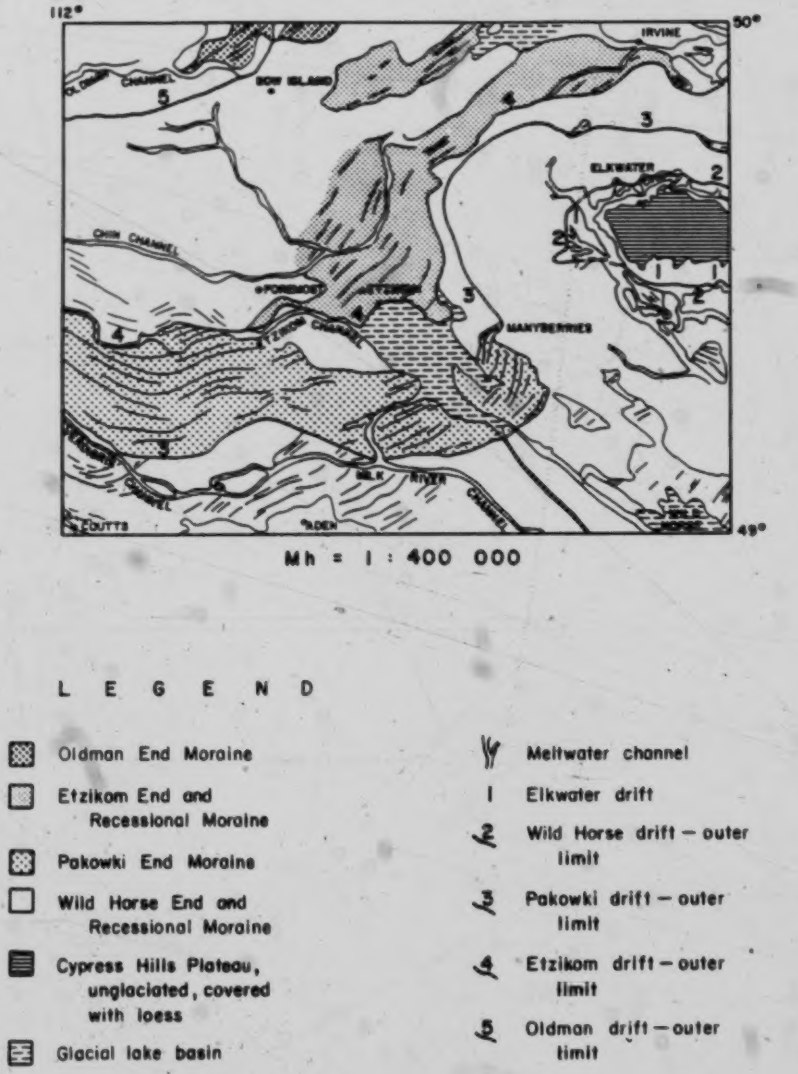

112°
50°
49°
IRVINE
OLDMAN CHANNEL
BOW ISLAND
ELKWATER
CHIN CHANNEL
FOREMOST
ETZIKOM
ETZIKOM CHANNEL
MANYBERRIES
MILK RIVER CHANNEL
COUTTS
ADEN
WILD HORSE
1
2
3
4
5
Mh = 1 : 400 000
L E G E N D
Oldman End Moraine
Etzikom End and Recessional Moraine
Pakowki End Moraine
Wild Horse End and Recessional Moraine
Cypress Hills Plateau, unglaciated, covered with loess
Glacial lake basin
Meltwater channel
1 Elkwater drift
2 Wild Horse drift — outer limit
3 Pakowki drift — outer limit
4 Etzikom drift — outer limit
5 Oldman drift — outer limit

Figure 5. Morphostratigraphic units in Foremost-Cypress Hills area.

SOURCE – ( after Westgate, 1968 )

similarity. Moreover, a high proportion of the sandstone is fine-grained. Thus, hydraulic conductivity values vary between $2.8 \times 10^{-9}$ and $2.8 \times 10^{-10}$ for the low, and between $2.8 \times 10^{-7}$ and $2.8 \times 10^{-8}$ m/s for the high permeability formations, respectively (Table 1). The bedrock slopes due north (Figure 6), and two buried valleys appear to exist within the basin.

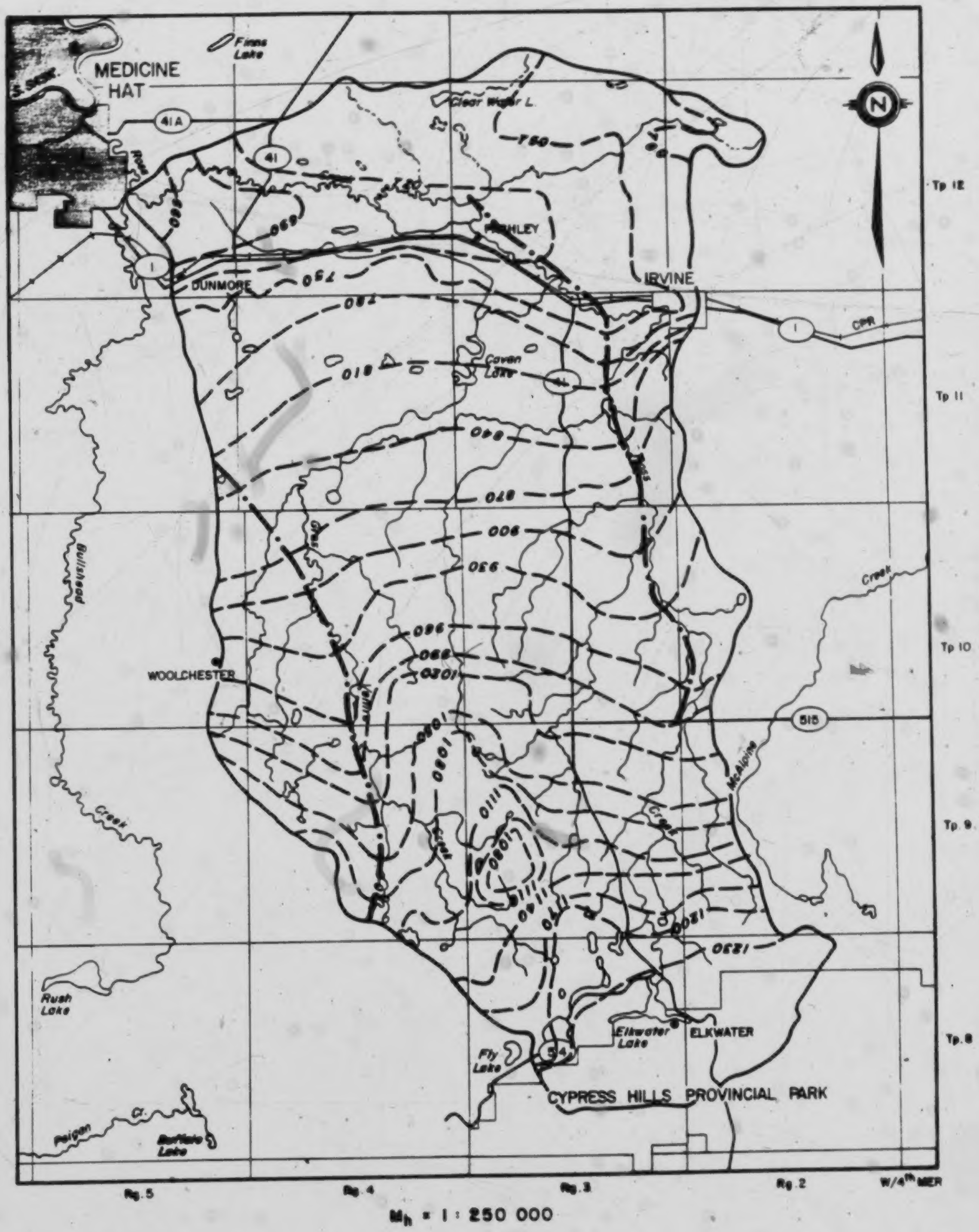
The glacial drift which varies from 0 to 30 m in thickness (Figure 7), is thickest at and in the vicinity of stream channels, and thinnest or non-existent on the Cypress Hills and in the northern plains of the basin. The drift sediments are lenticular in structure, and no extended continuous layers of any specific lithology are known to exist presently. Till is the main component of the glacial drift, whereas glacio-fluvial sand and gravel of meltwater channels, and glacio-lacustrine sand, silt and clay of preglacial and superglacial lakes are encountered commonly. Whereas the hydraulic conductivity of the till is about $2.8 \times 10^{-10}$ m/s, the local sand and gravel pockets may possess hydraulic conductivities of up to $5.6 \times 10^{-4}$ m/s (Table 1).

Both primary and secondary porosities are reported to control the occurrence of water in geologic materials of western Canada. The unconsolidated surficial deposits are dominated by primary porosity, although the abundance of joints in these sediments (Stalker, 1980) could be expected to influence their permeability significantly. Fractures and joints resulting from post-depositional tectonism, and effects of glaciation cause the development of secondary porosity in bedrock units. The resultant effective porosity of a unit is controlled by the magnitude of secondary porosity, abundance of clay, and the degree of cementation. Fracturing is usually conspicuous on outcrops of rock units, and drillers often log aquifers as "fractured shale" or "fractured sandstone". Evidence from well logs do not relate the basin's stratigraphy to fracture patterns or porosity. Although the materials in the basin may be assumed to be uniformly porous, no measured porosity data exist for either the surficial deposits or bedrock. Nevertheless, 22 to 35 percent has been estimated for similar materials in Alberta (Table 2). It is presumed that these porosities can be used for the geologic materials in Ross Creek Basin without severe errors.

Table 1. Hydraulic conductivity values of commonly occurring formations in Alberta.

| Formations | (K) Hydraulic conductivity (m/s) | | |
|---|---|---|---|
| | Probable range within formation | Probable range of average | Estimated average |
| Till, Lacustrine deposits, Lea Park, Pakowki, Alberta Gp., Wapiabi, Blackstone, Fernie Gp., Colorado Gp, Joli Fou, Labiche, Clearwater, Smoky Gp., Puskwaskau, Kaskapau, Shaftesbury, Low Spirit River, Loon River. | up to $5.6 \times 10^{-9}$ | up to $5.6 \times 10^{-10}$ | up to $2.8 \times 10^{-10}$ |
| Dunvegan, Pouce Coupe, Bearpaw, Aeolian deposits, Cardium, Blairmore Gp., Upper Manville, Top Spirit River. | $5.6 \times 10^{-11}$ to $5.6 \times 10^{-8}$ | $5.6 \times 10^{-10}$ to $5.6 \times 10^{-9}$ | $2.8 \times 10^{-9}$ |
| Edmonton, St. Mary, Wapiti, Belly River, Foremost, Oldman, Bulmonk, Willow Creek, Lower-Upper Manville. | $5.6 \times 10^{-10}$ to $5.6 \times 10^{-7}$ | $5.6 \times 10^{-9}$ to $5.6 \times 10^{-8}$ | $2.8 \times 10^{-8}$ |
| Milk River, Ribstone Creek, Lower Birch Lake, Bow Island, Vicking, Top Upper Maville, Bad Heart, Bluesby, Nikanamin. | $5.6 \times 10^{-9}$ to $5.6 \times 10^{-6}$ | $5.6 \times 10^{-8}$ to $5.6 \times 10^{-7}$ | $2.8 \times 10^{-7}$ |
| Paskapoo, Lower Manville, Ellerlie, Deville, Grand Rapids, Bullhead Gp., Codomin, Gelling, Pelican, Top Nikanamin. | $5.6 \times 10^{-8}$ to $5.6 \times 10^{-5}$ | $5.6 \times 10^{-7}$ to $5.6 \times 10^{-6}$ | $2.8 \times 10^{-6}$ |
| McMurray, Porcupine Hills. | | $5.6 \times 10^{-6}$ to $5.6 \times 10^{-5}$ | $2.8 \times 10^{-5}$ |
| Recent alluvial terraces, sands and gravells of buried valleys, sands and gravells of glacial meltwaters. | $5.6 \times 10^{-6}$ to $5.6 \times 10^{-3}$ | $2.8 \times 10^{-4}$ to $2.8 \times 10^{-3}$ | $5.6 \times 10^{-4}$ |

Source - Tóth, personal communication


MEDICINE HAT
S. Sask. R.
Finns Lake
Clear Water L.
41A
41
1
PASHLEY
DUNMORE
IRVINE
CPR
Cavan Lake
WOOLCHESTER
Bullshead
Ross
Creek
McAlpine
515
Rush Lake
Pelgan Cr.
Buffalo Lake
Fly Lake
Elkwater Lake
ELKWATER
CYPRESS HILLS PROVINCIAL PARK
750
780
810
840
870
900
930
960
990
1020
1050
1080
1110
1140
1170
1200
1230
Tp 12
Tp 11
Tp 10
Tp 9
Tp 8
Rg. 5
Rg. 4
Rg. 3
Rg. 2
W/4th MER
Mh = 1 : 250 000

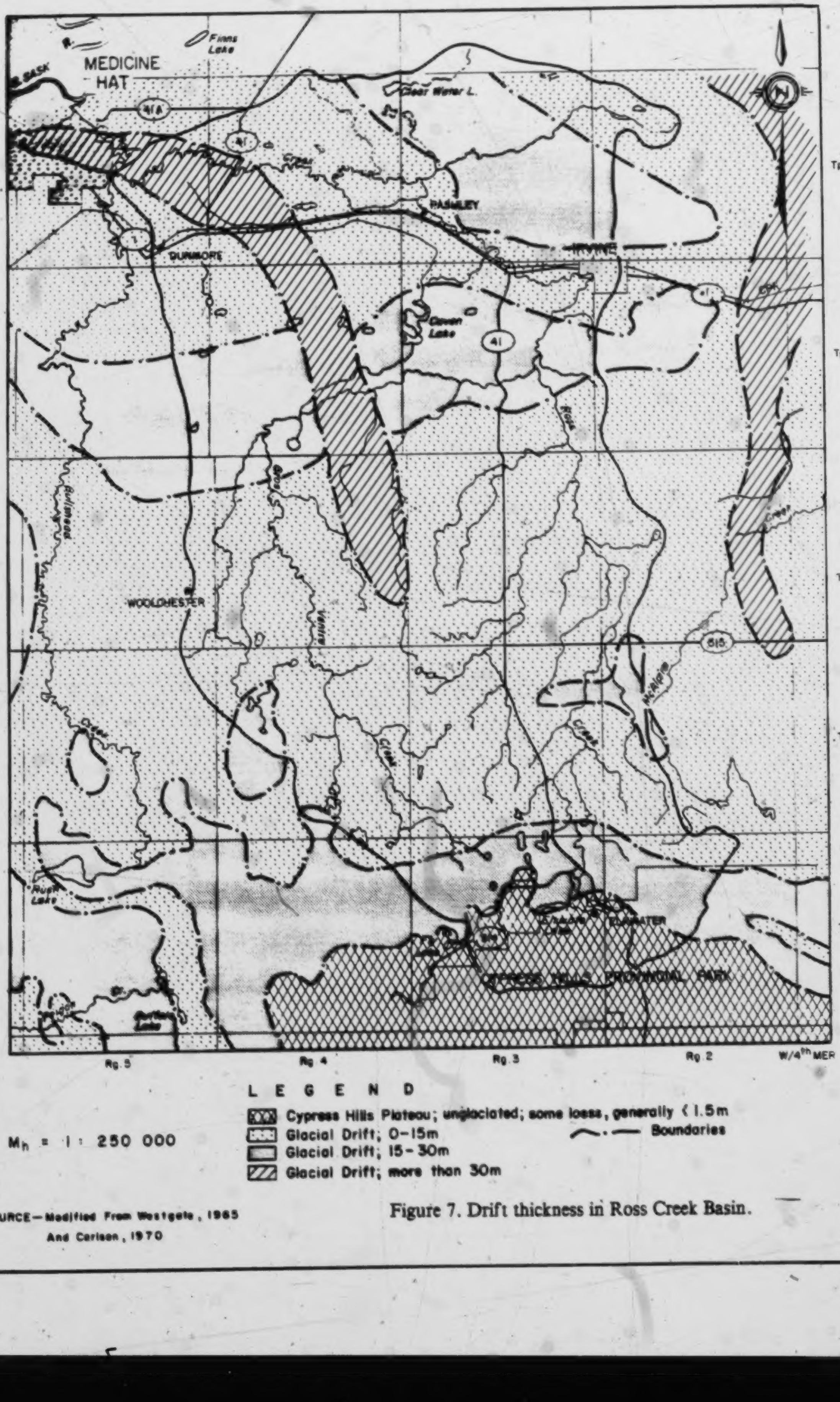
LEGEND

— —810— — Bedrock Contour Line

—.—.— Possible Buried Bedrock Valleys

SOURCE – Modified From Westgate, 1965
And Carlson, 1970

Figure 6. Bedrock topography of Ross Creek Basin.


MEDICINE HAT
Finns Lake
S. SASK. R.
41A
41
Clear Water L.
DUNMORE
IRVINE
Cavan Lake
Ross
WOOLCHESTER
Creek
515
CYPRESS HILLS PROVINCIAL PARK
ELKWATER
Rush Lake
Buffalo Lake
Tp 12
Tp 11
Tp 10
Tp 9
Tp 8
Rg. 5
Rg. 4
Rg. 3
Rg. 2
W/4th MER
LEGEND
Cypress Hills Plateau; unglaciated; some loess, generally < 1.5m
Glacial Drift; 0-15m
Glacial Drift; 15-30m
Glacial Drift; more than 30m
Boundaries
Mh = 1 : 250 000

SOURCE—Modified From Westgate, 1965
And Carlson, 1970

Figure 7. Drift thickness in Ross Creek Basin.

Table 2. Porosity values of some formations in Alberta.

| Formation | Lithology | Porosity | | |
|---|---|---|---|---|
| | | Type | Range % | Average % |
| Wapiti | sandstone + shale | intergranular | 34.2 - 11.2 | 22.0 |
| Milk River | | intergranular | 10.0 - ? | 10.0 |
| Alluvium River Bank | | intergranular | 35.0 - ? | 35.0 |
| Ellerslie | dolomite | vuggy + fracture | 30.0 - 12.5 | 20.0 |
| Nisku | dolomite | vuggy + fracture | 10.9 - 3.0 | 5.3 |
| Leduc | dolomite | vuggy + fracture | 10.8 - 3.0 | 7.1 |
| Cardium | sandstone | | 18.3 - 7.8 | 12.7 |
| Sunburst | sandstone | | 23.5 - ? | 23.5 |
| Wainwright | sandstone | intergranular | 31.0 - 25.0 | 26.7 |
| Cessford | | | 27.1 - 23.0 | 24.5 |
| Vicking | | vuggy | 23.0 - 7.9 | 15.4 |
| Fish Sc. | sandstone | | 23.0 - ? | 23.0 |
| Wabamun | | | 3.5 - 3.3 | 3.4 |
| Colony | | | 7.2 - ? | 7.2 |
| Swan Hills | | | 9.0 - 6.6 | 8.6 |
| Conrad | sandstone | | 17.4 - ? | 17.4 |
| Rundle | | | 10.0 - 7.0 | 8.5 |
| Banff | | | 11.0 - ? | 11.0 |
| Glauc. | sandstone | | 17.0 - 14.9 | 15.9 |
| Gilwood | sandstone | | 14.3 - ? | 14.3 |
| Elkton | | | 13.8 - 5.5 | 9.9 |
| Pekisko | | | 13.3 - ? | 13.3 |
| Ostracod | | | 20.8 - ? | 20.8 |
| Granite wash | | | 13.0 - 11.5 | 12.2 |
| Belly River | | | 19.0 - 15.0 | 17.3 |
| Arcs | | | 15.0 - 8.0 | 11.5 |
| Moulton | sandstone | | 17.5 - ? | 17.5 |
| Cooking Lake | | | 5.0 - ? | 5.0 |
| Todd-gray | | | 15.0 - ? | 15.0 |
| Cutbank | sandstone | | 21.0 - 17.3 | 19.2 |
| Turner River | | | 12.8 - 9.8 | 11.3 |

Source - Tóth, personal communication.

## 2.3 Hydrometeorology

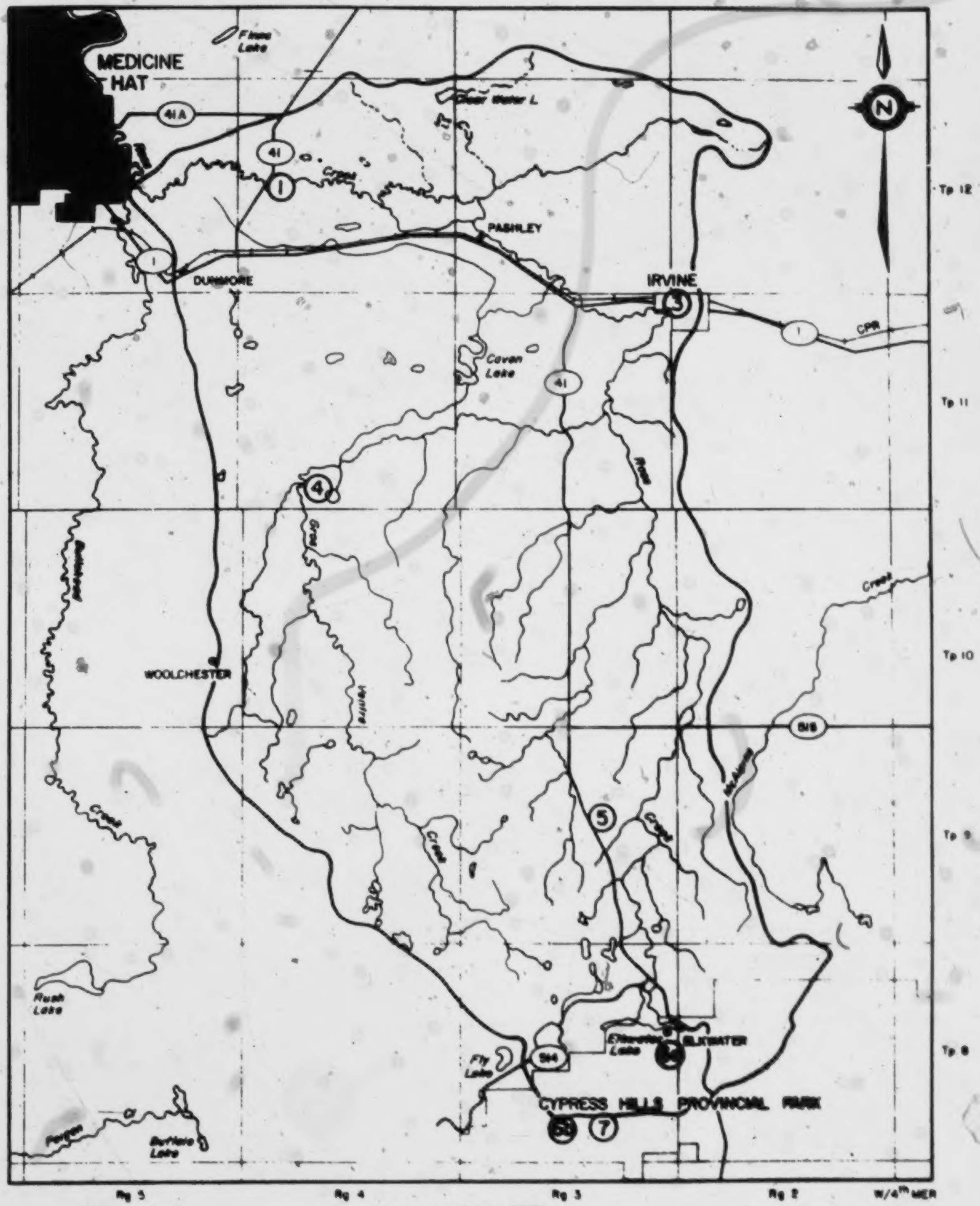
### 2.3.1 Precipitation and temperature

There are no meteorological stations within Ross Creek Basin with sufficiently long and complete records for analysis. However, a few precipitation gauging stations now exist in the basin (Figure 8), some of which were set up temporarily during the present study. Despite local climatic variations within the basin and its environs, records in Medicine Hat, the nearest station with reliable climatological data are used mainly in the hydrometeorological evaluation of the basin. These records are considered to simulate the meteorological conditions in the basin fairly, since Northwest Hydraulic Consultants Ltd. (1981b) correlated precipitation data in Medicine Hat with streamflow in Ross Creek successfully, at a correlation coefficient of 0.8.

Tables 3 and 4 list the monthly precipitation and mean monthly temperature in Medicine Hat (station 2) for the period of 1971 to 1980. The mean annual precipitation for this period is 327 mm. This average is close to the 1968 to 1971 average of 265 mm estimated by Wong (1974) in this area. Precipitation records for 1985, and for all the stations in the basin are reported in Ophori and Tóth (1985). Of these records, only those of stations 2, 6a and 7 are complete, and have annual totals of 278 mm, 408 mm and 496 mm, respectively. Using proportional calculations, stations 6a and 7 would produce mean annual precipitations of 479 mm and 583 mm, respectively, for the period for 1971 to 1980. Although these precipitation values are possibly higher than those expected at the northern end of the basin, they represent actual values within the basin in the area where most of the groundwater recharge is expected to occur. Thus, an average precipitation of 531 mm is estimated for the basin.

### 2.3.2 Evapotranspiration

The potential evapotranspiration is defined as the water which would be lost from a homogeneous cover of vegetation that never suffers from water shortage. Potential evapotranspiration exceeds actual evapotranspiration by a moisture deficit, the quantitative amount by which available moisture falls short of the climatic demand for water. Water surplus is the amount of moisture lost from the soil by subsurface runoff, after the soil has been recharged to capacity, if precipitation exceeds potential evapotranspiration. Water surplus is, therefore, a crude estimate of the amount of water available for surface runoff and


MEDICINE HAT
Fleas Lake
Clear Water L.
41A
41
1
Creek
PASHLEY
DUNMORE
IRVINE
3
CPR
Caven Lake
4
Ross
Tp 12
Tp 11
Tp 10
Tp 9
Tp 8
WOOLCHESTER
Creek
Creek
515
5
Creek
Rush Lake
Fly Lake
514
Elkwater Lake
ELKWATER
CYPRESS HILLS PROVINCIAL PARK
7
Buffalo Lake
Rg 5
Rg 4
Rg 3
Rg 2
W/4th MER

$M_h$ = 1 : 250 000

L E G E N D

① STATION NUMBER (Alberta Environment, Lethbridge)

STATION 2 is at Medicine Hat Airport

Figure 8. Precipitation stations in Ross Creek Basin.

Table 3. Monthly precipitation data in Medicine Hat (mm).

| Year | Jan | Feb | Mar | Apr | May | Jun | Jul | Aug | Sep | Oct | Nov | Dec | Annual total |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 1980 | 23.4 | 8.0 | 13.7 | 10.8 | 42.0 | 64.7 | 28.4 | 36.4 | 23.4 | 37.6 | 5.1 | 23.2 | 316.7 |
| 1979 | 5.7 | 16.3 | 8.0 | 26.1 | 45.9 | 76.3 | 32.4 | 20.4 | 14.9 | 18.4 | 6.7 | 10.8 | 281.9 |
| 1978 | 18.9 | 16.2 | 8.7 | 72.4 | 55.4 | 53.4 | 49.9 | 33.9 | 92.6 | 31.4 | 22.6 | 8.5 | 463.9 |
| 1977 | 18.0 | 0.7 | 6.2 | 6.7 | 80.5 | 19.9 | 41.2 | 40.2 | 39.3 | 1.8 | 17.9 | 31.0 | 303.4 |
| 1976 | 3.0 | 9.3 | 20.9 | 26.5 | 22.9 | 58.7 | 86.7 | 40.8 | 4.8 | 6.8 | 14.1 | 6.3 | 300.8 |
| 1975 | 7.6 | 20.4 | 36.3 | 52.9 | 86.2 | 69.6 | 36.8 | 34.0 | 22.7 | 14.4 | 25.7 | 25.5 | 432.1 |
| 1974 | 14.4 | 17.4 | 27.5 | 44.4 | 52.7 | 44.9 | 22.9 | 47.9 | 3.3 | 5.0 | 1.5 | 7.6 | 289.5 |
| 1973 | 1.8 | 8.6 | 0.8 | 30.5 | 5.8 | 123.0 | 6.6 | 37.0 | 19.4 | 17.4 | 23.7 | 18.9 | 293.5 |
| 1972 | 21.7 | 23.2 | 8.3 | 19.4 | 39.3 | 41.1 | 43.9 | 33.0 | 47.6 | 20.0 | 4.3 | 29.7 | 331.5 |
| 1971 | 42.3 | 7.3 | 16.1 | 9.3 | 45.4 | 40.8 | 9.3 | 20.7 | 26.7 | 18.1 | 9.1 | 12.6 | 257.7 |
|  |  |  |  |  |  |  | Mean annual |  |  |  |  |  | 327.1 |

Source - Atmospheric Environment Service (AES)

Table 4. Mean monthly temperature in Medicine Hat (°C).

| Year | Jan | Feb | Mar | Apr | May | Jun | Jul | Aug | Sep | Oct | Nov | Dec |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 1980 | -12.5 | -7.0 | -2.8 | 11.2 | 14.7 | 17.0 | 19.4 | 16.2 | 13.3 | 8.2 | 2.6 | -9.2 |
| 1979 | -18.2 | -16.7 | 0.4 | 3.4 | 10.6 | 16.9 | 20.4 | 20.1 | 16.1 | 9.3 | -1.3 | -2.0 |
| 1978 | -16.9 | -13.1 | -2.4 | 6.3 | 12.1 | 17.1 | 19.5 | 17.6 | 14.0 | 8.5 | -6.0 | -10.8 |
| 1977 | -12.0 | 0.9 | 1.4 | 9.3 | 12.9 | 18.6 | 19.1 | 16.0 | 12.0 | 7.8 | -3.5 | -12.7 |
| 1976 | -7.5 | -2.1 | -2.7 | 8.7 | 14.2 | 15.2 | 20.1 | 19.9 | 15.6 | 5.9 | -1.3 | -4.6 |
| 1975 | -8.4 | -14.3 | -5.9 | -0.5 | 10.9 | 15.1 | 21.7 | 16.2 | 13.1 | 6.5 | -2.4 | -7.4 |
| 1974 | -12.5 | -4.0 | -3.8 | 8.0 | 9.6 | 18.7 | 20.6 | 16.3 | 11.8 | 9.2 | 0.5 | -2.3 |
| 1973 | -6.3 | -6.9 | 3.3 | 4.1 | 13.9 | 16.6 | 19.8 | 19.7 | 12.9 | 8.1 | -8.7 | -6.8 |
| 1972 | -16.2 | -12.9 | -1.1 | 5.6 | 12.9 | 18.0 | 17.3 | 21.0 | 9.5 | 4.0 | -0.1 | -14.0 |
| 1971 | -14.7 | -5.8 | -2.9 | 6.8 | 13.2 | 16.1 | 18.9 | 23.0 | 11.5 | 5.9 | -1.2 | -11.5 |

Source - Atmospheric Environment Service (AES) groundwater recharge, an important aspect of water balance studies.

### 2.3.3 Soil water balance

The soil water balance of an area is an account of all water gains, losses, and storages in the soils of that area over a period of time. In it, potential evapotranspiration, actual evapotranspiration, water deficit, water surplus, and soil moisture storage which are significant in the understanding and use of the water resources in an area, are computed.

In this study, the empirical methods of Thornthwaite (1948), Thornthwaite and Mather (1955), and Mather (1978) have been applied in the evaluation of the water balance parameters from the Medicine Hat data. Although the reliability of the Thornthwaite's mean temperature method is questionable when applied to short periods of time (Pelton et al., 1960), the method is often found to estimate evapotranspiration and moisture surplus in the right order of magnitude. The analysis was carried out for five different soil moisture capacities varying from 0.5 mm to 250 mm, and covering the range estimated by the Department of Agriculture, Canada (1956) as:

Sandy loam: 107 mm
Loam and silty loam: 158 mm
Clayloam and silty clay loam: 183 mm
Clay and silty clay: 198 mm
Heavy clay: 224 mm

The results of the analysis for the various soil types are presented in Appendix 2. Water balance graphs have been drawn for a dry year (1973), a ten-year average (1971 to 1980), and a wet year (1978), using the results and an arbitrarily selected moisture capacity of 50 mm (Figure 9). The graphs depict the relationships between the two largest components of the water system, precipitation and evapotranspiration in Ross Creek Basin.

Figure 9 shows that precipitation in the area is nonuniform and occurs above 20 mm during seven months of the year. Precipitation averages 58 mm in June which appears to be the wettest month on the average, whereas the driest month may be November or February with an average precipitation of 13 mm. Actual evapotranspiration is virtually nonexistent in mid-October, thereafter, it decreases gradually to zero in November in response to falling temperatures and inactive vegetation, resulting in a much higher range than that of

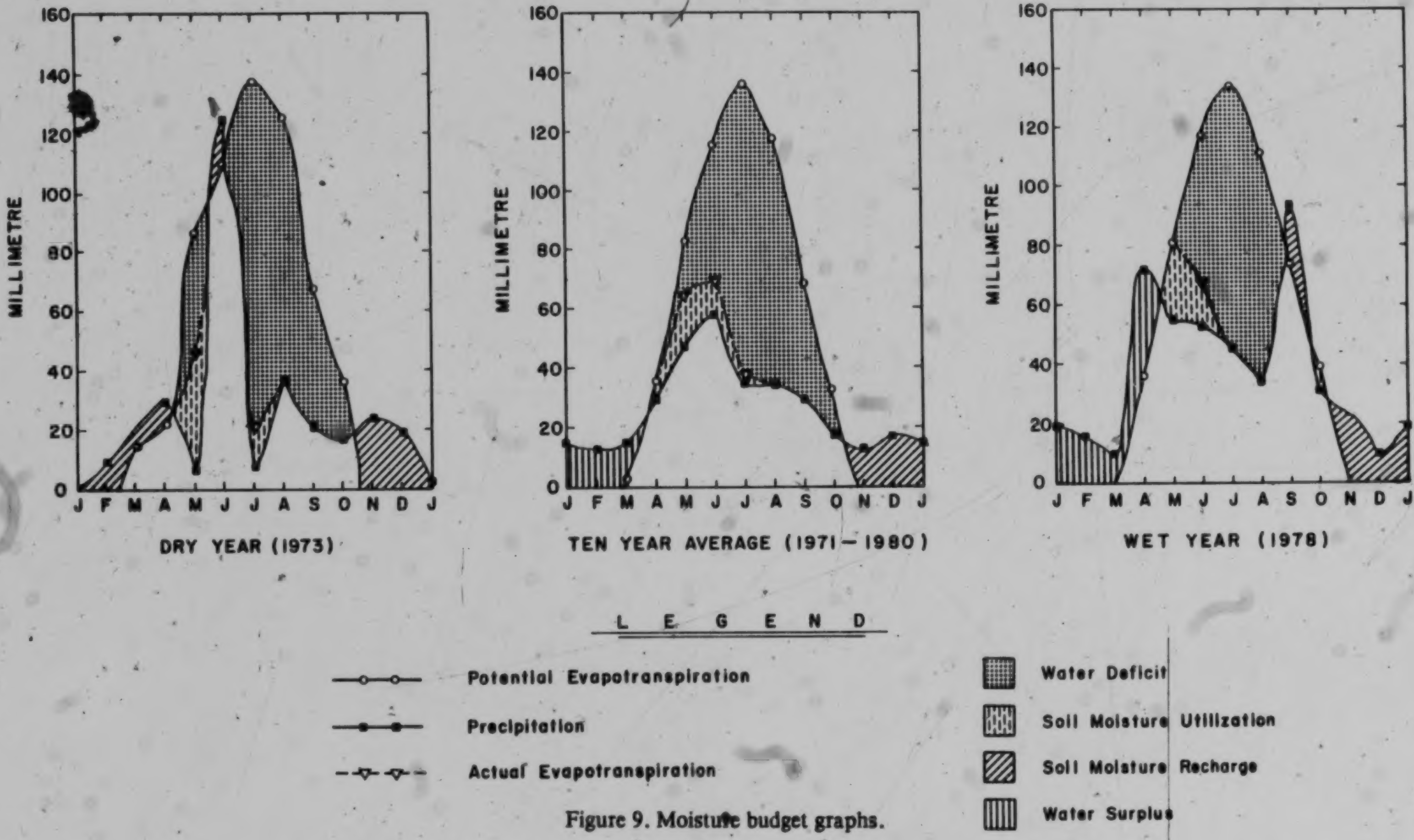


Figure 9. Moisture budget graphs.

precipitation.

On the average, precipitation is greater than potential evapotranspiration during five months of the year (November to March), but in the other seven months, potential evapotranspiration exceeds precipitation, and reaches a maximum of 146 mm in July. The three periods of the annual water cycle experienced commonly in various parts of North America are only observable during average conditions in the study area: a period of soil moisture utilization which is accompanied usually by soil moisture deficit, since available soil moisture is incapable of balancing the difference between precipitation and potential evapotranspiration; a period of soil moisture recharge; and finally, a period of water surplus during which excess water is available for subsurface runoff and groundwater recharge. The period of soil moisture utilization occurs between April and August. On the other hand, soil moisture recharge occurs between October and January, but it could be suspended during the frost season when precipitation exists mainly in the form of snow. On melting, the snow, and increasing rates of precipitation result in moisture surplus between February and April.

The three water balance graphs (Figure 9), may be compared as follows: during the dry year, there was no period of water surplus, and a prolonged period of water deficit existed between April and October. The ten-year average graph had a short period of moisture surplus (February to April), and a similar extended period of moisture deficit as that of the dry year occurred between May and October. This situation suggests that the average hydrometeorological conditions in the study area represent a true average between dry and wet years. During the wet year (1978), the period of water surplus was extensive, from February to May, whereas the period of moisture deficit was relatively short, covering the months of May to September only.

In Table 5, several parameters are compared for the dry, average, and wet year water balances. A moisture deficit is depicted by the table in all three water balance conditions. The moisture deficits in all three conditions are dissimilar, being highest in the dry year, intermediate in the average condition, and lowest in the wet year as expected. Whereas moisture surplus is 79 mm in the wet year, no moisture surplus is recorded for the dry year, and only 35 mm is recorded for average conditions. It is remarkable that no moisture surplus exists in a dry year, and the moisture surpluses of the average and wet years may also be less if a soil moisture capacity of 100 mm is used in these calculations. An annual moisture surplus of zero was

Table 5. Calculated moisture budget parameters.

| Condition | Precipitation (mm) | Evapotranspiration | | Moisture deficit (mm) | Moisture surplus (mm) |
|---|---|---|---|---|---|
| | | Potential (mm) | Actual (mm) | | |
| Dry year (1973) | 279 | 589 | 289 | 309 | 0 |
| Ten year average (1971-1980) | 323 | 602 | 312 | 275 | 35 |
| Wet year (1976) | 459 | 592 | 376 | 216 | 79 |

estimated also by Wong (1974) for the period between 1968 and 1971.

It is also observed from Table 5 that 289 mm (79 percent) of the annual precipitation of 297 mm was lost through evapotranspiration in the dry year (1973), leaving a balance of 8 mm. During the ten-year average conditions, total precipitation was 323 mm of which 312 mm or 96 percent was returned to the atmosphere by actual evapotranspiration, resulting in a balance of 11 mm. In the wet year (1978), precipitation was 459 mm, with an evapotranspiration value of 376 mm (82 percent), and a left over of 83 mm. There is evidence therefore, that after evapotranspiration losses, residual precipitation ranges between 8 mm and 83 mm per year, with an average of 11 mm. Accompanying the annual variation are seasonal and daily variations, which all combine to give a picture of a continuously fluctuating water system in the area of study. Such a variable water system is not a consistent source of water supply without a well planned management program.

### 2.4 Mapping of field manifestations of groundwater flow

The consideration of the relationship between hydraulic discharge of groundwater and certain naturally occurring field phenomena in regional groundwater studies is on the rise (Meinzer, 1923, 1927; Meyboom 1966; Tóth, 1966a). Based on field mapping alone, Tóth (1966a) separated groundwater discharge zones from midline and recharge zones in an area around Trochu, Alberta. Since then, several studies have indicated the applicability and usefulness of field hydrogeological mapping in groundwater studies (eg. Clissold, 1967; Lissey, 1968; Freeze, 1969a; Tóth, 1971; Leskiw, 1971).

Field mapping techniques are particularly useful for the following reasons: 1. in undeveloped areas where information about the groundwater regime is scarce or nonexistent, and where climatic conditions are in favour of the proposed type of investigation, they become the sole source of data from which groundwater resources may be evaluated; 2. they form the basis for reconnaissance of the extent, location, and distribution of groundwater flow systems, a useful guide in the planning of the economical location and spacing of test holes for the purposes of water-supply development programs; 3. they may help in solving specific groundwater problems typical of the Plains Regions, such as finding suitable locations for dug-out type water supplies, drainage sloughs, obtaining a preliminary idea of the possible consequences of irrigation, and explaining the development of certain soil types; 4. they employ field phenomena indicative of near-surface groundwater conditions, which are usually easy to identify and to interpret; and 5. they are much less costly in reconnaissance-type studies than test drilling. For some of the reasons mentioned above, the mapping techniques played a useful role in the present study. Another unique aspect of the techniques herein, is their applicability to Ross Creek Basin, a typically semi arid part of the Plains Regions. This situation tends to suggest that, after all, field mapping techniques may be applicable in all the commonly encountered climatic types in the Plains Regions of Alberta.

Field mapping was carried out during the months of June, July and August, 1983. Traverses were planned systematically in order to visit every quarter section in the basin, in search of hydrogeological information, as well as to use roads as much as possible. In planning the traverses, extensive use was made of air photographs, and of existing data and reports. In particular, the information obtained from the files of the Groundwater Information Section of Alberta Environment, from Alberta Hydrogeological Information Map Series, and from Alberta Hydrogeological Reconnaissance Map Series of the Research Council of Alberta were helpful. In the field, each point of observation containing important groundwater information was assigned a number, and the physical environment, topography, kind of vegetation, quality of crops, moisture conditions, and water levels in wells were noted. The presence or absence of salt affected soils was also noted. Water samples for chemical analysis were collected from bodies of surface water, springs, water wells, and occasionally, field analysis for major element concentrations was done with the aid of a field Hach Kit. The conductivity and temperature of the water samples were also measured. Where it was impossible to conduct actual mesurements in the field, data on well depths, depths to water, well lithologies, and well and spring productivity history were either obtained from local people or deduced from previous reports.

A classification of observed phenomena helps in their collection and interpretation. Tóth (1966a) classified the phenomena found in the Trochu area into two main groups. His classification appears to cover the observed phenomena in Ross Creek Basin. The details of the classification, and the method of interpretation of field phenomena are presented in Tóth (1966a). The observed features in the present study are listed below, and are displayed in Figure 10:

1. springs;
2. seepages;

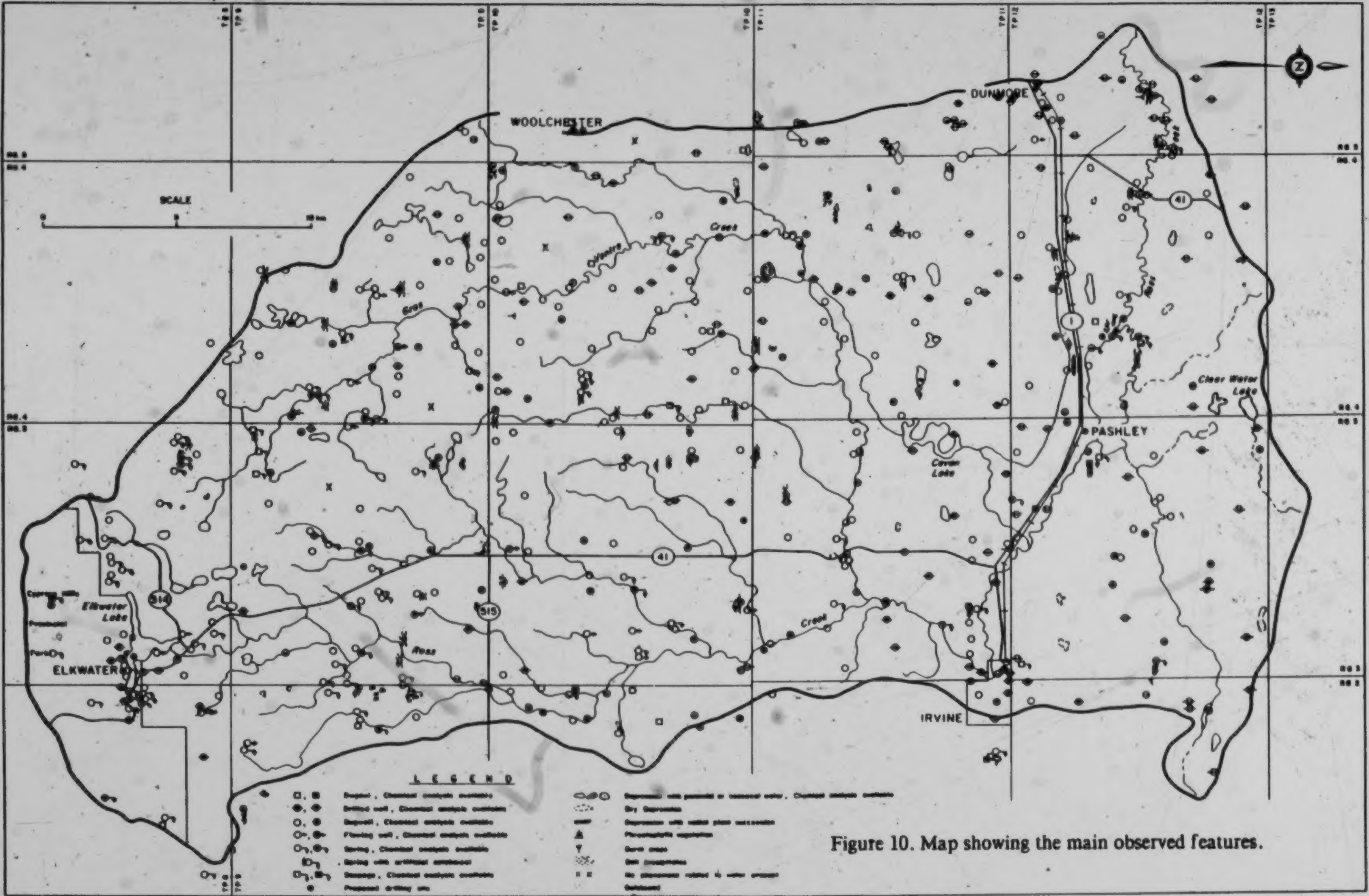

WOOLCHESTER
DUNMORE
PASHLEY
IRVINE
ELKWATER
Elkwater Lake
Clear Water Lake
Cavan Lake
Creek
Ross
41
1
515
514
SCALE
0
10 km
LEGEND
N

Figure 10. Map showing the main observed features.

3. groundwater levels;
4. flowing wells;
5. chemical quality of water (distribution of chemical components);
6. physical quality of water;
7. natural vegetation;
8. salt precipitates;
9. burnt crops;
10. quicksand;
11. moist depressions;
12. dry depressions;
13. man-made objects.

A detailed description of the observed features is given in Ophori and Tóth (1983); only the results are summarized here. For convenience in reference to features, observation point numbers are indicated in Figure 11.

One of the major objectives of the field mapping was to estimate the groundwater flow pattern in the basin, as much as observable surficial manifestations of groundwater would permit. In order to achieve this objective, the simple relationship between the observed phenomena described earlier and the groundwater regime in the study area must be understood thoroughly.

The field observed features are produced by either surface water, groundwater, or both. It is apparent therefore, that areas in which groundwater moves toward the land surface will possess more moisture than can be accounted for by local climatic conditions alone, whereas, in areas where groundwater moves away from the land surface and is constantly being replenished by surface water, a deficiency in moisture content above the saturated zone will be registered. Consequently, areas of upward-moving groundwater (discharge areas) are characterized by discharge features, such as, springs, seepages, shallow water tables, flowing wells, high electrical conductivity waters, high total dissolved solids, phreatophytes, salt precipitates, burnt crops, quicksand, moist depressions, and some types of man-made objects. Often, however, not all the features listed may be found at one observation point. Areas of downward-moving groundwater (recharge areas), on the other hand, are indicated by moisture deficient features, such as, dry depressions, moist depressions containing water of relatively low electrical

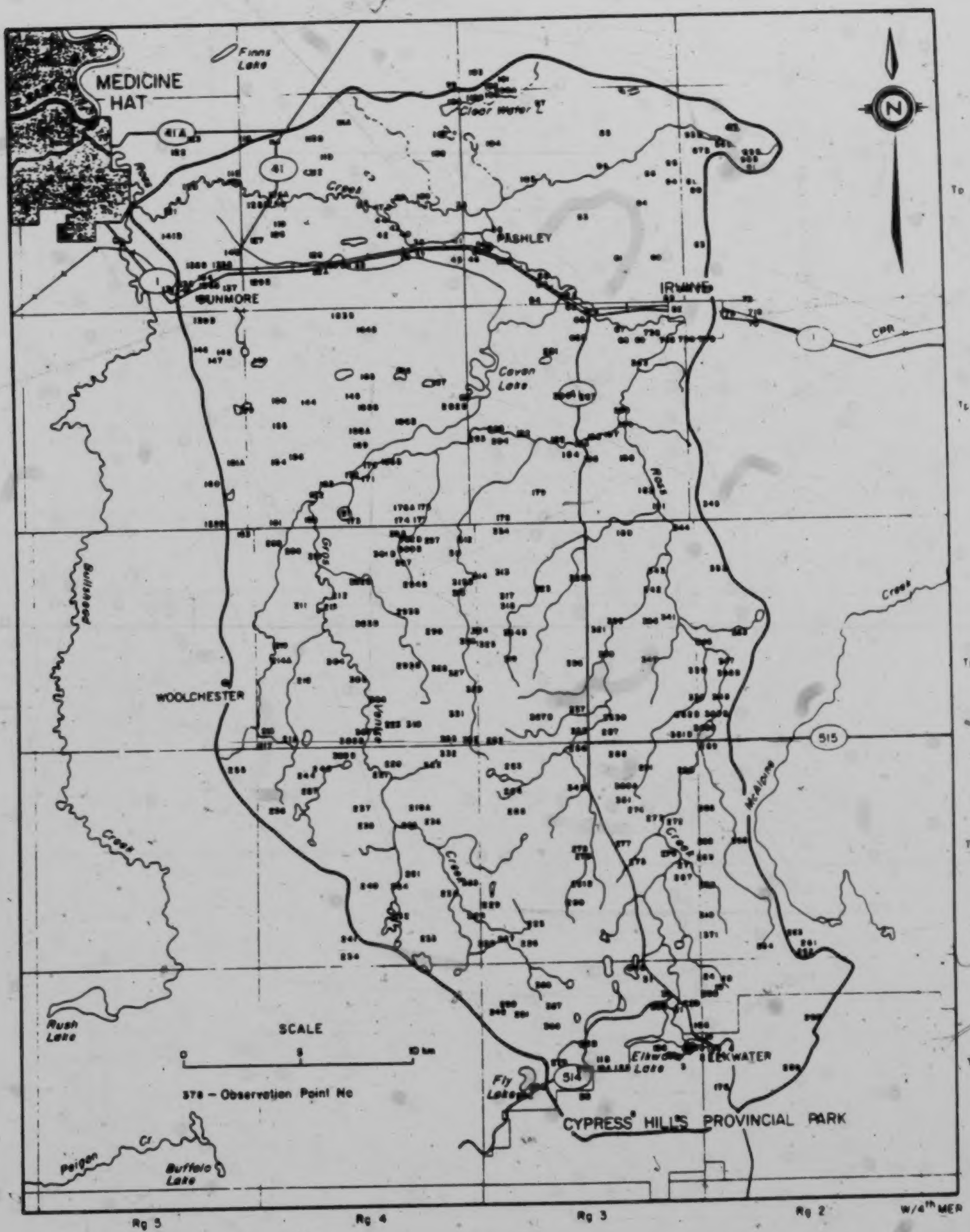

Finns Lake
MEDICINE HAT
41A
41
Clear Water L
Ross
Creek
ASHLEY
IRVINE
1
DUNMORE
CPR
Cavan Lake
Ross
Bullshead
Gros Ventre
WOOLCHESTER
Creek
515
McAlpine
Creek
Creek
Rush Lake
SCALE
0
5
10 km
378 – Observation Point No
Fly Lake
514
Elkwater Lake
ELKWATER
CYPRESS HILLS PROVINCIAL PARK
Peigan Cr
Buffalo Lake
Rg 5
Rg 4
Rg 3
Rg 2
W/4th MER
Tp 12
Tp 11
Tp 10
Tp 9
Tp 8
N

Figure 11. Map of observation points.

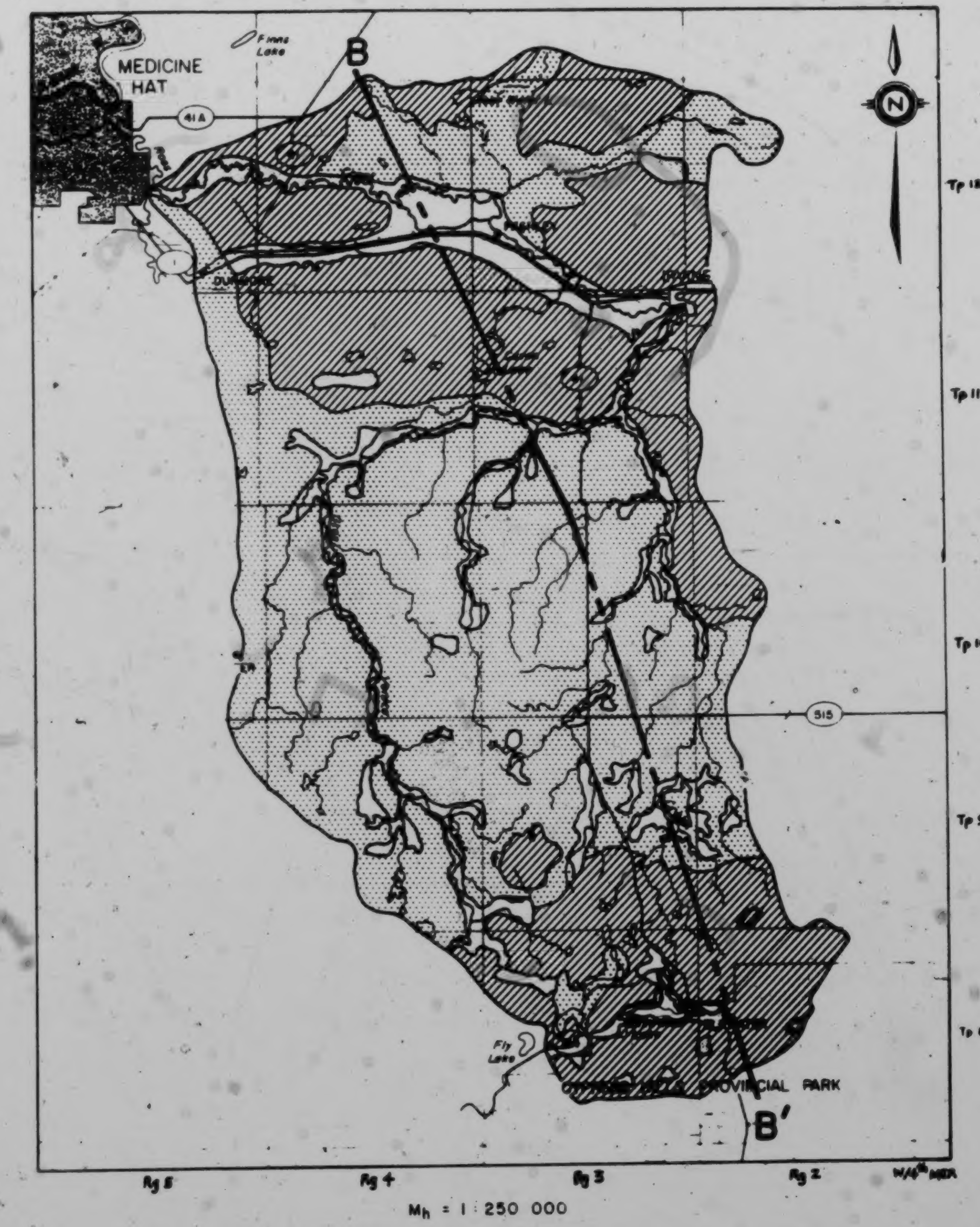
conductivity, low total dissolved solids, phreatophytes (may or may not be present), relatively deep water tables, and the lack of typical discharge features (springs, seepages, flowing wells, salt precipitates, burnt crops, quicksand). As nature is not always clear cut, discharge and recharge phenomena may either co-exist, or may not occur in some areas. Such areas are mapped as uncertain.

The criteria outlined above have been applied, along with the field observations (Figure 10), in the construction of Figure 12, the final product of the mapping. Figure 12 shows that water moving away from the regional high areas around the Cypress Hills is discharged in the regional low areas near the Trans-Canada Highway, after part has been discharged at the intermediate topographic elevations. The figure also suggests that most of the discharge occurs along, but is not restricted to the Ross and Gros Ventre Creeks, and their tributaries. Whereas the local recharge areas on both sides of the lower Ross Creek, and to the east-central corner of the map-area are presumably contributing to discharge along the creek, it has not been possible on the basis of field phenomena to outline the local recharge areas possibly feeding the isolated, minor discharge areas. The areas within the basin, where observed features do not favour either discharge or recharge conditions, or where such typical features have not been identified, have been mapped as uncertain in Figure 12. These zones may correspond to the midline areas in which groundwater flow is rectilinear and nearly horizontal. Based on the results of field mapping alone, a generalized schematic groundwater flow pattern for the basin is shown in Figure 13.

Using Figure 12 and a rough estimate with a planimeter, the map-area is covered by 24 percent discharge area, 35 percent recharge area, and 41 percent midline area, indicating that discharge is not confined to the thalwegs of the channels.

Leskiw (1971), and Maclean (1974) have pointed out the relationship between soil types and groundwater movement in central Alberta. In Ross Creek Basin, this relationship is observable. A comparison of Figures 12 and 14 shows discharge areas to be associated with clay, eroded and river bottom soils; midline areas with silty loam; and recharge areas with loam, fine sandy loam, fine sand, and mixed soils. Any disparity in this conclusion may be related to the time difference in the construction of the two figures.

Whereas most of the springs in the basin discharge and loose all their water in their immediate vicinities, groundwater discharge from few springs is significant and contribute to


MEDICINE HAT
Finns Lake
B
41A
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DUNMORE
IRVINE
N
Tp 12
Tp 11
Tp 10
Tp 9
Tp 8
515
Fly Lake
CYPRESS HILLS PROVINCIAL PARK
B′
Rg 5
Rg 4
Rg 3
Rg 2
W/4th MER
Mh = 1 : 250 000
LEGEND
▨ Recharge area
☐ Discharge area
Uncertain area
B-B′ TRACE OF SECTION IN FIGURE 10

Figure 12. Distribution of recharge and discharge areas.

SOURCE-Adapted From Ophori & Toth,1983

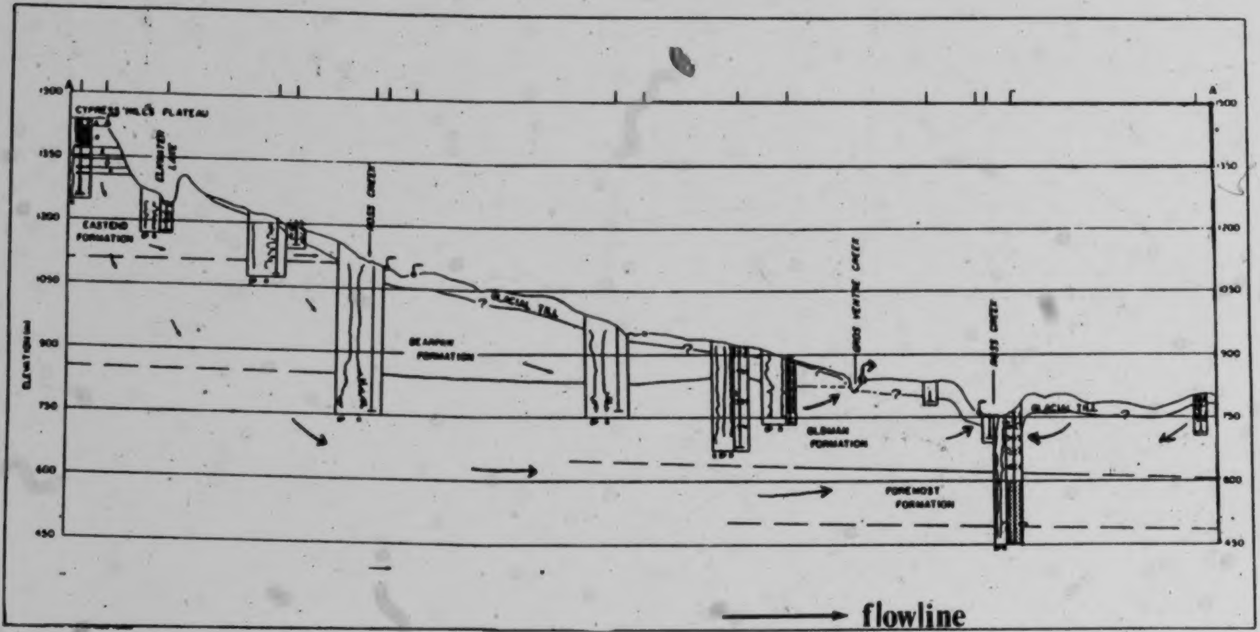
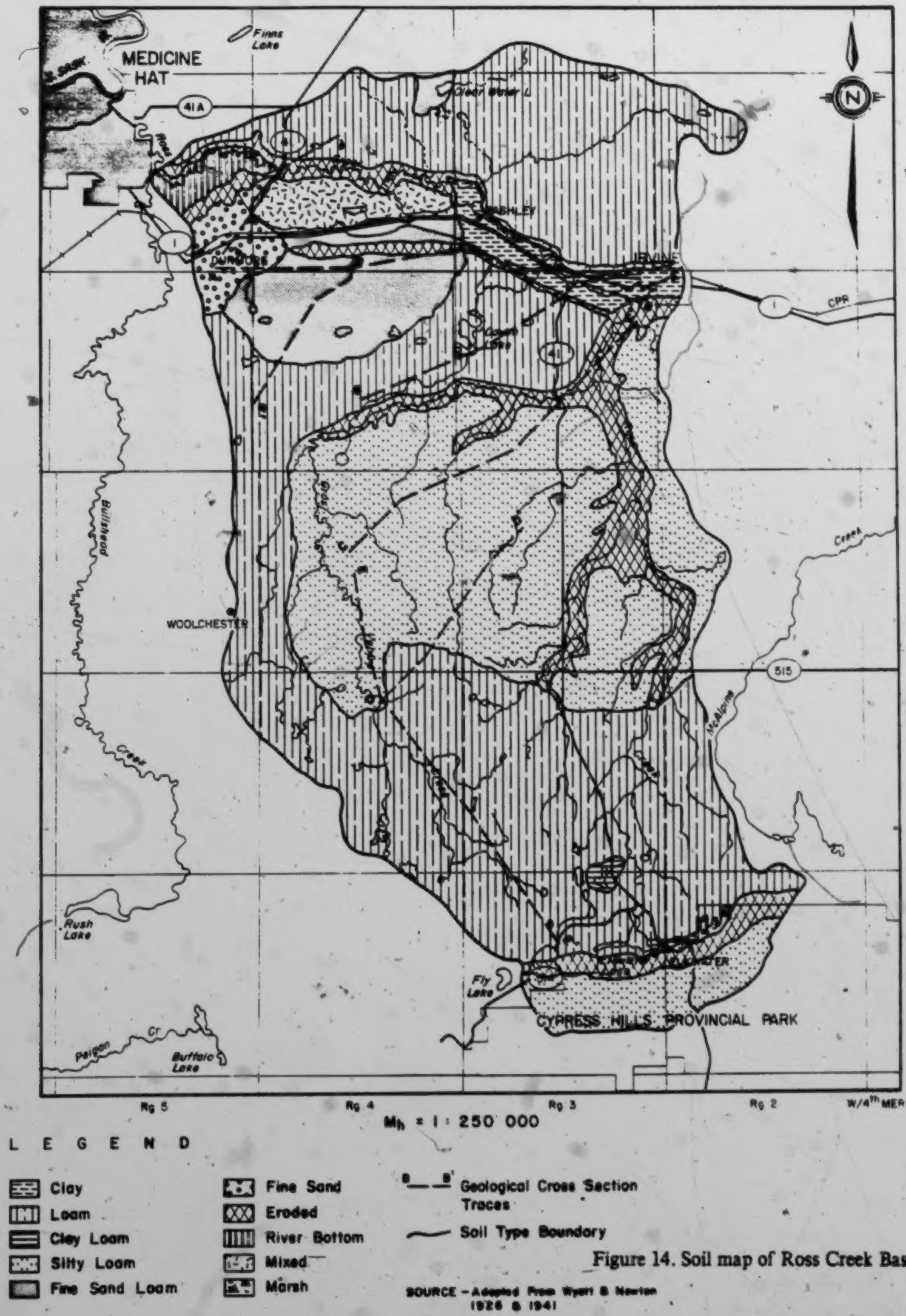

A
A
ELEVATION (m)
1500
1350
1200
1050
900
750
600
450
CYPRESS HILLS PLATEAU
ELEVATED LAKE
EASTEND FORMATION
ROSS CREEK
GLACIAL TILL
BEARPAW FORMATION
GROS VENTRE CREEK
OLDMAN FORMATION
ROSS CREEK
GLACIAL TILL
FOREMOST FORMATION
?
flowline

Figure 13. Schematic groundwater flow pattern based on field mapping.


MEDICINE HAT
S. SASK.
Finns Lake
41A
Clear Water L.
Ross
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Bullshead
Creek
WOOLCHESTER
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McAlpine
Creek
Rush Lake
Fly Lake
ELKWATER
CYPRESS HILLS PROVINCIAL PARK
Peigan Cr
Buffalo Lake
Rg 5
Rg 4
Rg 3
Rg 2
W/4th MER
$M_h$ = 1 : 250 000
LEGEND
Clay
Loam
Clay Loam
Silty Loam
Fine Sand Loam
Fine Sand
Eroded
River Bottom
Mixed
Marsh
B — B' Geological Cross Section Traces
Soil Type Boundary

Figure 14. Soil map of Ross Creek Basin.

SOURCE - Adapted From Wyatt & Newton 1926 & 1941 the water resources of some local inhabitants downslope from the points of discharge. Local channels or coulees attend springs with significant discharge rates. The possible origin and evolution of tributary water courses on broad hillsides, where a continuous outflow of groundwater coincides with soft and permeable pockets of surface materials is indicated in Tóth (1966a). The tributary channels associated with the springs (8, 9, 176, 259) near the top of the Cypress Hills, and which form the headwaters of Elkwater Lake, appear to have originated from the down-cutting effects of discharge from the springs. These relatively large springs presumably issue their water out of the Frenchman-Ravenscrag Formation (Cretaceous-Tertiary) contact. The springs at observation points 40, 41 and 44 probably formed the tributary channel to Ross Creek at that location, and has been dredged artificially into a perennial stream by the landowner.

### 2.5 Aquifer hydraulics

Several hydraulic parameters are tied with groundwater hydrology, and function as the basis for aquifer identification. The most common of these parameters are porosity (n), transmissivity (T), hydraulic conductivity (K), and storage coefficient (S). In particular, once the transmissivity is known, estimates of groundwater yield are made empirically.

### 2.5.1 Collection of data

In Alberta, the past few decades have been a period of prolific water well drilling that has generated a significant amount of shallow, semi-quantitative hydrogeologic data. After completion of a well, drilling contractors usually conduct a short bail test or pumping test to determine well yield. For a domestic well, the period of bailing or pumping is commonly about 30 minutes, and then the total drawdown is noted. Time-drawdown measurements are rarely taken, and observation wells are non-existent. Occasionally, time-recovery data are recorded, and few pumping tests of several hours duration are performed. This information has been of value to several groundwater studies in Alberta (Tóth, 1966b; Garven, 1980), at the reconnaissance stage.

In Ross Creek Basin, this approach appears to be of little value for the following reasons: 1. such data have been recorded for only few sites in the basin; 2. aquifers anticipated on the basis of geology (Milk River and Oldman-Foremost Formations) occur at greater depths than depths for which such data are recorded; 3. the lengths of many of these bail and pumping tests, and their accuracy in measurement is probably questionable as reporting proper aquifer testing procedure. In any event, all available data were analyzed in order to establish initial estimates of the local hydraulic characteristics of the subsurface units. In addition to these data, various tests were conducted during and after the drilling program in the present study. Five sites were drilled on the basis of results from the field mapping and a knowledge of the geology. Compressor tests were conducted during the drilling. Two pumping and recovery tests were performed at the most promising sites identified during the drilling. Finally, several slug (piezometer) tests were conducted at the drilled sites. All the test data may be found in Ophori and Tóth (1983, 1985).

### 2.5.2 Analysis of data

The compressor tests were conducted as part of the well completion procedure. In these tests, air is forced with an air compressor attached to the drill rig into the open hole to wash out the hole. As the air is forced into the hole at a controlled rate, water is "blown" out of the hole at a rate that can be related to the yield of the hole. The yield is then estimated by measuring the rate of flow of water out of the hole. This method, though approximate, produces yields that can be considered as reliable first approximation.

Piezometer tests are tests carried out in single piezometers to determine in situ hydraulic conductivity values. The tests are inititated by causing an instantaneous change in the water level in piezometers through a sudden introduction (slug test) or removal (bail test) of a known volume of water. The recovery of water with time is then measured. In the present study, the sudden change in water level was caused by introducing a solid cylinder of known volume. Freeze and Cherry (1979) describe the method of interpretation of the resulting water level versus time data. Using this method, the equation for the hydraulic conductivity is (Hvorslev, 1951):

$$K=\frac{r^2 \ln(L/R)}{2LT_0} \qquad (1)$$

where: K = hydraulic conductivity [L/T];

r = radius of piezometer [L];

R = radius of intake pipe [L];

L = length of intake pipe [L];

$T_0$ = basic time lag [T].

For each unit, a transmissivity value was obtained by multiplying the hydraulic conductivity by the thickness of the unit. Once the transmissivity was obtained, a yield value was estimated through the equation (Tóth, 1966b):

$$Q_{20} = \frac{TH \times 0.7}{2110} \quad (2)$$

where: $Q_{20}$ = 20 year safe yield (igpm);

$T$ = transmissivity (igpd/ft);

$H$ = available drawdown (ft).

$Q_{20}$ is the constant rate at which a well can produce continuously so that at the end of twenty years, the water level will be drawn down to the top of the producing aquifer.

For the tests in which total drawdowns only were recorded, rough estimates of transmissivity through the calculation of "apparent transmissivity" (Farvolden, 1961) were made. Specifically, this method is suitable for evaluating specific capacity. The method, which under or over estimates the actual transmissivity, depending whether the drawdown curve flattens or steepens with time respectively, employs the modified non-equilibrium formula of Cooper and Jacob (1946). It assumes a straight line plot between 0.1 minutes and the time of the single observation of total drawdown.

The water level measurements obtained from pumping and recovery tests were evaluated according to the "straight-line" or Jacob's method (Cooper and Jacob, 1946). Since the measured points fell on straight lines on semilog-arithmetic plots, transmissivity values were calculated from the data obtained from the discharge wells without the use of observation wells. The equation for estimating the transmissivity is (Freeze and Cherry, 1979):

$$T = \frac{0.18Q}{\Delta h} \quad (3)$$

where: $Q$ = well discharge $[L^3/T]$;

$T$ = transmissivity $[L^3/T]$;

$\Delta h$ = drawdown per log cycle [L].

In three cases where there were data from an observation well, transmissivity was obtained from a logarithmic plot of the time-drawdown data (Theis, 1935). In these cases storage coefficient values of 0.001, 0.001 and 0.002 were obtained. For the data and graphs used in these calculations, reference should be made to Ophori and Tóth (1983, 1985).

### 2.5.3. Results

The calculated hydraulic conductivity, transmissivity, and yield values are shown in Table 6. From these values, transmissivity and groundwater yield maps have been constructed as in Figures 15 and 16, respectively.

It is clear from Figure 15 that there are no extensive bodies of high transmissivity (greater than $1.0\times10^{-4}$ m$^3$/s) in the basin. Only few local pockets of geologic materials which may be termed as aquifers on this basis occur. However, in combining transmissivity with available drawdown, most of these pockets become insignificant in yielding groundwater (Figure 16). Strictly speaking, Ross Creek Basin possesses a poor potential to yield groundwater from storage. There are only two localities in which groundwater can be produced at rates of between $7.6\times10^{-4}$ and $4.1\times10^{-3}$ m$^3$/s to cause complete mining of the resource in twenty years. One of these localities lies in the property of the Cypress Hills Provincial Park.

### 2.6 Hydrogeologic units

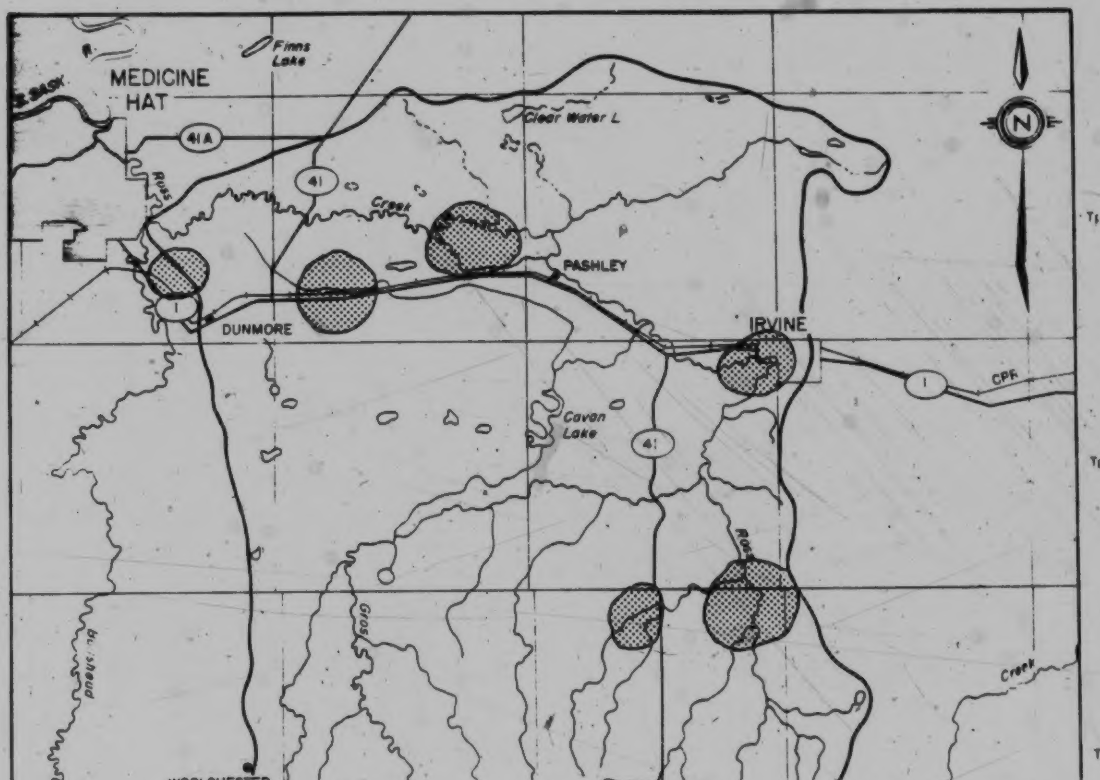
A hydrogeologic unit is a body of rock which is uniform hydraulically, independent of the geologic boundaries. The rocks in Ross Creek Basin may be grouped into two hydraulically distinct sets of units: those which are capable of producing groundwater at rates in excess of $7.6\times10^{-4}$ m$^3$/s, and those which produce less (Figure 16). These two categories of rock units are referred to as aquifers and aquitards in the remaining sections of this report. Details regarding this classification may be obtained from Ophori (1986).

Table 6. Calculated hydraulic conductivity, transmissivity and yield values.

| Values from pevious data | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|
| Location | Type of data | Type of test | Method | Transmissivity (m²/s) | Location | Type of data | Type of test | Method | Transmissivity (m²/s) |
| NW-19-08-2W4 | DD | PW | Jacob | 5.2 × 10⁻⁴ | NW-25-11-5W4 | - | - | AP | 3.4 × 10⁻⁵ |
| NW-19-08-2W4 | DD | PW | Jacob | 6.5 × 10⁻⁴ | NW-25-11-5W4 | - | - | AP | 4.2 × 10⁻⁴ |
| NW-19-08-2W4 | RE | PW | Theis | 5.6 × 10⁻⁴ | NW-26-11-5W4 | - | - | AP | 1.7 × 10⁻⁵ |
| NW-19-08-2W4 | RE | PW | Theis | 6.3 × 10⁻⁴ | NW-26-11-5W4 | - | - | AP | 4.9 × 10⁻⁵ |
| 23/24-08-3W4 | DD | PW | Jacob | 4.7 × 10⁻⁴ | NW-26-11-5W4 | - | - | AP | 7.1 × 10⁻⁵ |
| -24-08-3W4 | DD | PW | Jacob | 4.7 × 10⁻⁴ | NW-26-11-5W4 | - | - | AP | 6.6 × 10⁻⁵ |
| NE-36-11-3W4 | DD | PW | Jacob | 2.4 × 10⁻⁴ | NW-26-11-5W4 | - | - | AP | 6.9 × 10⁻⁶ |
| NE-36-11-3W4 | RE | PW | Theis | 2.1 × 10⁻⁴ | NW-34-11-5W4 | - | - | AP | 8.8 × 10⁻⁵ |
| NE-36-11-3W4 | DD | OW | Jacob | 2.8 × 10⁻⁴ | NW-35-11-5W4 | - | - | AP | 8.2 × 10⁻⁶ |
| NE-36-11-3W4 | RE | OW | Theis | 2.7 × 10⁻⁴ | SE-01-12-3W4 | - | - | AP | 7.0 × 10⁻⁵ |
| NE-36-11-3W4 | DD | OW | Theis | 1.4 × 10⁻³ | SE-04-12-3W4 | - | - | AP | 6.1 × 10⁻⁶ |
| SE-24-08-3W4 | - | - | AP | 5.2 × 10⁻⁴ | NW-06--2-3W4 | - | - | AP | 6.6 × 10⁻⁵ |
| SE-24-08-3W4 | - | - | AP | 8.1 × 10⁻⁶ | SW-15-12-3W4 | - | - | AP | 1.5 × 10⁻⁵ |
| SE-24-08-3W4 | - | - | AP | 1.2 × 10⁻³ | SE-20-12-3W4 | - | - | AP | 1.9 × 10⁻⁵ |
| 15-24-08-3W4 | - | - | AP | 1.8 × 10⁻⁴ | NE-27-12-3W4 | - | - | AP | 6.8 × 10⁻⁵ |
| 16-24-08-3W4 | - | - | AP | 1.4 × 10⁻⁴ | SE-30-12-3W4 | - | - | AP | 2.5 × 10⁻⁴ |
| SW-02-09-3W4 | - | - | AP | 4.4 × 10⁻⁶ | SW-33-12-3W4 | - | - | AP | 9.6 × 10⁻⁶ |
| NE-16-09-4W4 | - | - | AP | 3.3 × 10⁻⁵ | SE-04-12-4W4 | - | - | AP | 1.9 × 10⁻⁵ |
| SE-15-10-3W4 | - | - | AP | 1.6 × 10⁻⁵ | SW-08-11-3W4 | - | - | AP | 2.7 × 10⁻⁴ |
| NW-36-10-3W4 | - | - | AP | 3.1 × 10⁻⁴ | NE-13-12-4W4 | - | - | AP | 9.4 × 10⁻⁶ |
| NE-21-10-3W4 | - | - | AP | 7.1 × 10⁻⁶ | SE-10-12-4W4 | - | - | AP | 3.6 × 10⁻⁵ |
| SW-28-10-4W4 | - | - | AP | 4.2 × 10⁻⁵ | SE-10-12-4W4 | - | - | AP | 1.3 × 10⁻⁵ |
| SW-28-10-4W4 | - | - | AP | 3.0 × 10⁻⁶ | SW-28-12-4W4 | - | - | AP | 5.5 × 10⁻⁵ |
| SW-28-10-4W4 | - | - | AP | 2.6 × 10⁻⁵ | SW-28-12-3W4 | - | - | AP | 2.6 × 10⁻⁵ |
| SW-16-11-3W4 | - | - | AP | 1.4 × 10⁻⁵ | NW-32-12-3W4 | - | - | AP | 2.3 × 10⁻⁶ |
| NE-19-11-3W4 | - | - | AP | 1.6 × 10⁻⁶ | NW-01-12-3W4 | - | - | AP | 1.1 × 10⁻⁶ |
| SW-23-11-3W4 | - | - | AP | 2.3 × 10⁻⁶ | SE-02-12-5W4 | - | - | AP | 5.1 × 10⁻⁵ |
| NW-30-11-3W4 | - | - | AP | 4.3 × 10⁻⁵ | NW-03-12-5W4 | - | - | AP | 7.0 × 10⁻⁶ |
| NE-30-11-3W4 | - | - | AP | 4.8 × 10⁻⁶ | NW-03-12-5W4 | - | - | AP | 4.3 × 10⁻⁵ |
| SE-33-11-3W4 | - | - | AP | 1.6 × 10⁻⁵ | NW-10-12-3W4 | - | - | AP | 5.0 × 10⁻⁴ |
| NE-36-11-3W4 | - | - | AP | 1.9 × 10⁻⁴ | NW-14-12-5W4 | - | - | AP | 8.9 × 10⁻⁶ |
| NE-36-11-3W4 | - | - | AP | 2.8 × 10⁻⁶ | SE-23-12-5W4 | - | - | AP | 1.1 × 10⁻⁵ |
| NE-36-11-3W4 | - | - | AP | 2.8 × 10⁻⁴ | SE-23-12-5W4 | - | - | AP | 8.9 × 10⁻⁶ |
| SW-06-11-4W4 | - | - | AP | 2.8 × 10⁻⁶ | NE-26-12-5W4 | - | - | AP | 6.0 × 10⁻⁵ |
| SW-26-11-4W4 | - | - | AP | 1.1 × 10⁻⁵ | SW-36-12-5W4 | - | - | AP | 2.8 × 10⁻⁶ |
| NE-12-11-3W4 | - | - | AP | 5.3 × 10⁻⁶ | -36-12-5W4 | - | - | AP | 4.0 × 10⁻⁶ |

| Values from present data | | | | | | | | |
|---|---|---|---|---|---|---|---|---|
| Location | Piezometer test | | | Pump test | | | Compressor test | |
| | Hydraulic cond. (m/s) | Transmissivity (m²/s) | +Yield (m³/s) | Hydraulic cond. (m/s) | Transmissivity (m²/s) | Yield (m³/s) | Hole no | Yield (m³/s) |
| 12-08-08-2W4 | 9.0 × 10⁻⁷ | 1.4 × 10⁻⁶ | 9.7 × 10⁻⁶ | | | | 2084E | 0 - 3.8 × 10⁻⁵ |
| 06-24-08-3W4 | 8.4 × 10⁻⁶ | 1.5 × 10⁻⁶ | 2.2 × 10⁻⁴ | 1.7 × 10⁻⁵<br>1.4 × 10⁻⁵ | 3.0 × 10⁻⁵<br>2.6 × 10⁻⁵ | 4.0 × 10⁻⁴ | 2093E | 7.6 × 10⁻⁵ |
| 08-33-10-3W4 | 1.2 × 10⁻⁶ | 4.9 × 10⁻⁶ | 4.1 × 10⁻⁵ | | | | 2079E | 1.5 × 10⁻⁶ - 2.3 × 10⁻⁶ |
| 16-25-11-4W4 | 1.2 × 10⁻⁶ | 6.3 × 10⁻⁶ | 4.4 × 10⁻⁶ | | | | 2080E | 3.8 × 10⁻⁵ - 7.6 × 10⁻⁵ |
| | | | | | | | 2081E<br>2082E | 0<br>3.8 × 10⁻⁵ - 7.6 × 10⁻⁵ |
| 01-14-12-4W4 | 4.2 × 10⁻⁶ | 5.1 × 10⁻⁶ | 2.1 × 10⁻⁵ | 5.2 × 10⁻⁷<br>4.6 × 10⁻⁶ | 6.3 × 10⁻⁶<br>6.6 × 10⁻⁶ | 2.5 × 10⁻⁶ | 2086E | 9.1 × 10⁻⁶ - 1.5 × 10⁻⁵ |
| | | | | | | | 2088E | 1.5 × 10⁻⁶ - 2.3 × 10⁻⁶ |

RE = Recovery; DD = Drawdown; AP = Apparent; OW = Observation well; PW = Pumping well

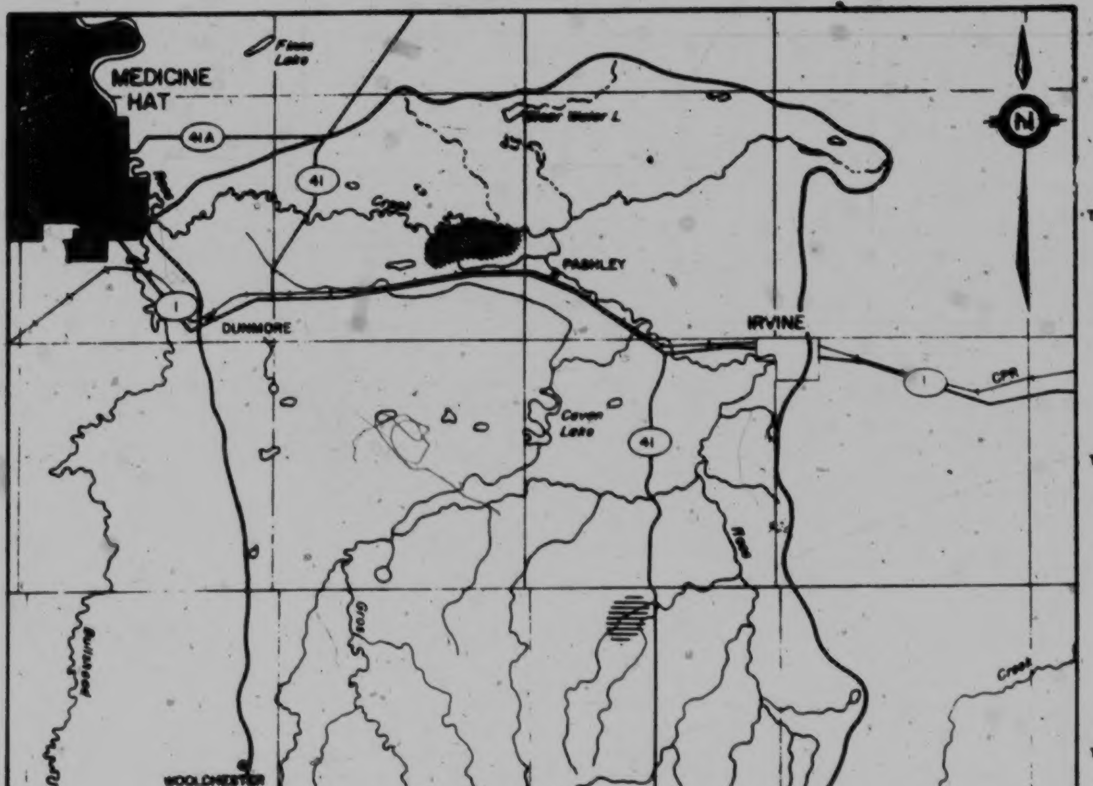

MEDICINE HAT
S. SASK.
Finns Lake
Clear Water L
41A
41
Ross
Creek
PASHLEY
DUNMORE
1
IRVINE
CPR
Cavan Lake
Gros
Ventre
WOOLCHESTER
Creek
McAlpine
515
Rush Lake
Fly Lake
514
ELKWATER
CYPRESS HILLS PROVINCIAL PARK
Pelgar Cr
Buffalo Lake
Tp 12
Tp 11
Tp 10
Tp 9
Tp 8
Rg 5
Rg 4
Rg 3
Rg 2
W/4TH MER
N
Mh = 1 : 250 000

L E G E N D

☐ Less Than $1.0 \times 10^{-4}$ m$^2$/sec

▨ Greater Than $1.0 \times 10^{-4}$ m$^2$/sec

Figure 15. Transmissivity distribution, Ross Creek Basin.


MEDICINE HAT
41A
41
Creek
Water L
PASHLEY
DUNMORE
IRVINE
1
CPR
Cavan Lake
41
Ross
Gros
Bullshead
WOOLCHESTER
515
Creek
Creek
Rush Lake
Fly Lake
514
CYPRESS HILLS PROVINCIAL PARK
Buffalo Lake
Tp. 12
Tp. 11
Tp. 10
Tp. 9
Tp. 8
Rg 5
Rg 4
Rg 3
Rg 2
W/4th MER
$M_h$ = 1 : 250 000

L E G E N D

| | |
|---|---|
| ☐ | 0-1·5 (chance possible of 1·5-2·3) ($\times 10^{-4}$ m$^3$/s) |
| ▤ | 1·5-2·3 ( " ) |
| ■ | 2·3-7·6 ( " ) |
| ■ | 7·6-41 ( " ) |

Figure 16. Groundwater yield map of Ross Creek Basin.

## 2.7 Hydrogeochemistry

### 2.7.1 Data collection and treatment

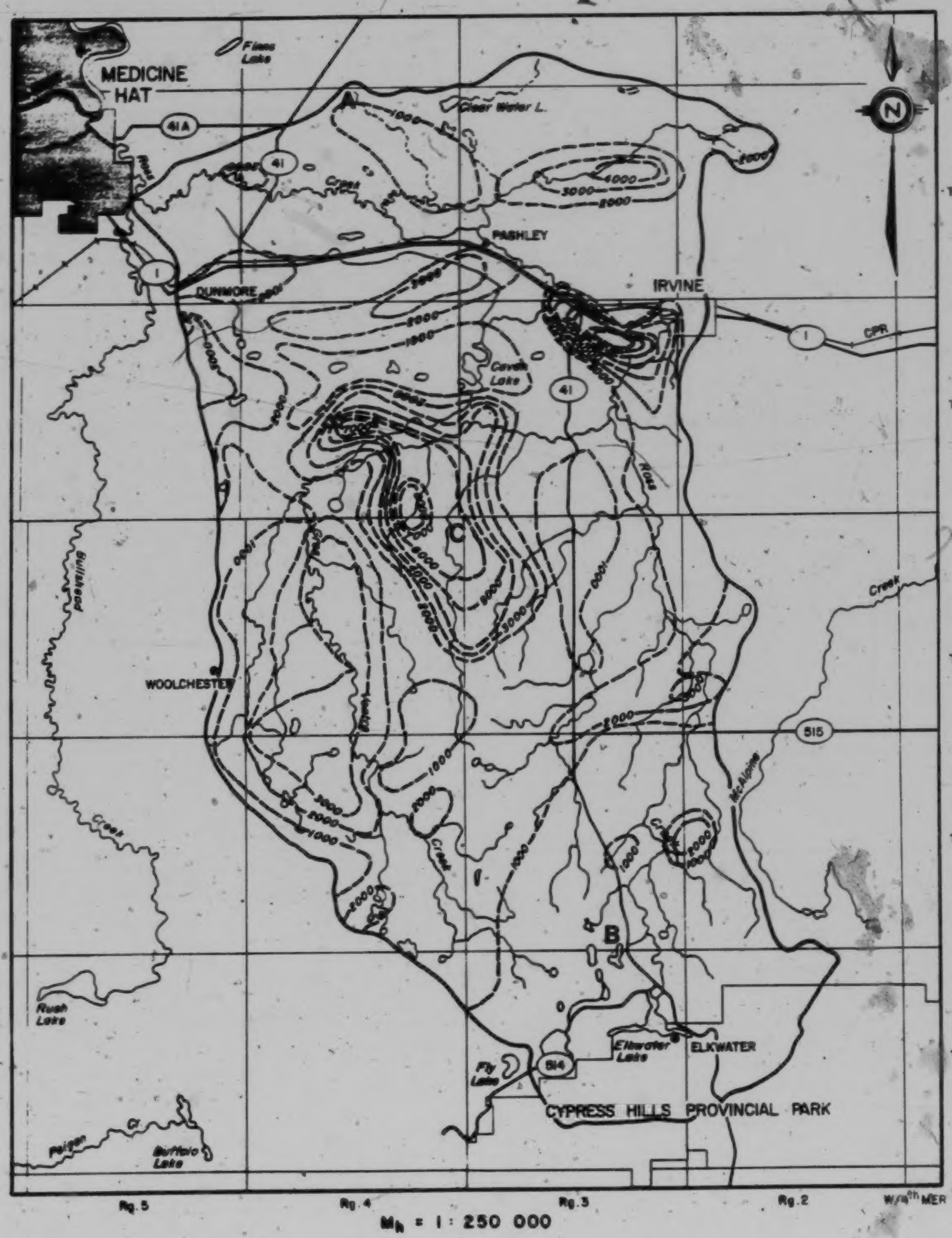
In order to determine the relationship between water chemistry and flow systems, and the potability of groundwater in the basin, a hydrogeochemical analysis of water samples was necessary.

Over 167 samples of well water, springs and surface water have been analyzed. Of these, 67 were pre-existing records obtained from the Groundwater Information Section of Alberta Environment, and 102 were analyses of water samples collected during the summer field mapping. In total, 149 samples received a complete or partial laboratory analysis which included determination of pH, electrical conductivity, sodium absorption ratio (SAR), total hardness, alkalinity, total dissolved solids, and ionic concentrations of all major anions and cations. Most of the laboratory analyses were carried out at the Water and Wastewater Laboratory of Alberta Environment. The remainder received a short field analysis performed with a portable Hach Kit and conductivity bridge, which included the determination of ionic concentrations of calcium, magnesium, chloride, sulphate, bicarbonate, iron, and the determination of pH and electrical conductivity.

### 2.7.2 Distribution of chemical constituents

Whereas all the results of the hydrogeochemical analyses are enclosed in Ophori and Tóth (1983, 1985), the distribution of some of the chemical constituents are displayed here in the form of contour maps (Figures 17 to 24).

Figure 17 illustrates the areal distribution of total dissolved solids of water wells and springs only, sloughs were not considered. A first glance at Figure 17 shows that groundwater in the basin is highly mineralized. On the average, about one-quarter of the basin is underlain by water with total dissolved solids greater than 2 000 ppm, whereas about one-half of the basin has total dissolved solids less than or equal to 1 000 ppm. Mineralization is least at the north and south drainage divides, and increases toward the lower elevations along the Ross Creek valley at the Trans-Canada Highway. However, this simple picture is complicated by locally high values of total dissolved solids, the peak of which occurs in the central part of the map-area.


MEDICINE HAT
Flaco Lake
Clear Water L.
41A
41
Crest
PASHLEY
DUNMORE
IRVINE
CPR
1
Caven Lake
Ross
Tp 12
Tp 11
Tp.10
Tp. 9
Tp. 8
C
B
WOOLCHESTER
Bullshead
Creek
515
McAlpine
Creek
Rush Lake
Fly Lake
514
Elkwater Lake
ELKWATER
CYPRESS HILLS PROVINCIAL PARK
Pelgan Cr.
Buffalo Lake
Rg. 5
Rg. 4
Rg. 3
Rg. 2
W/4th MER
1000
2000
3000
4000
$M_h$ = 1 : 250 000

L E G E N D

1000 Total Dissolved Solids Contour

Figure 17. Areal distribution of total dissolved solids.

The relative abundance of calcium to magnesium, expressed as the ratio $Ca^{2+}$ : $Mg^{2+}$ is believed to change only slightly along short paths of groundwater movement (Tóth, 1966b). Thus, the high variation in the observed ratio (Figure 18) does not necessarily indicate groundwater flow directions, but may reflect different original $Ca^{2+}$ : $Mg^{2+}$ ratios of rocks in the recharge areas.

The areal distribution of $HCO_3^- + CO_3^{2-}$ as a percentage of the total anions in ppm is shown in Figure 19. But for some scattered occurrences, $HCO_3^- + CO_3^{2-}$ content is relatively high (higher than 75%) in the vicinity of the Cypress Hills, and around and west of Cavan Lake. Generally, bicarbonates and carbonates constitute more than 50 percent of the anions in the southern one-third of the basin, and of the area immediately south of the Trans-Canada Highway. The percentage decreases away from these two areas toward the central part of the basin, and to some strips along the Trans-Canada Highway where the value falls short of 25 percent.

The distribution of $SO_4^{2-}$ is also presented as a percentage of the total anions in ppm (Figure 20). Sulphate is the next important major anion to bicarbonate and carbonate in the groundwater of the area of study. As depicted by Figure 20, sulphate content ranges from less than 20 percent to greater than 80 percent. The lowest values occur in the areas north of the Cypress Hills and south of the Trans-Canada Highway, from where they increase gradually toward the central portions of the map-area. Strips of high sulphate content (greater than 80%) occur also along the Trans-Canada Highway. A comparison of Figures 19 and 20 shows that areas of high sulphate content can be related to those of low bicarbonate-carbonate content, and vice versa, with some degree of certainty.

Except for seven local zones where concentration occurs above 100 ppm, chloride content is below 100 ppm in the map-sheet (Figure 21). The largest zone of relatively high chloride content is observed to coincide with peak values of sulphate, and with minimum values of bicarbonate and carbonate ions.

The average fluoride content is less than 0.5 mg/l, although a value of 6.0 mg/l was determined in one well at the east-central margin of the basin (Figure 22).

According to Figure 23, sodium and potassium content as percentages of total cations seldom occurs above 80 percent, and is usually less than 60 percent in the groundwater of more than half of the map-area. No particular trend is observed in the sodium and potassium

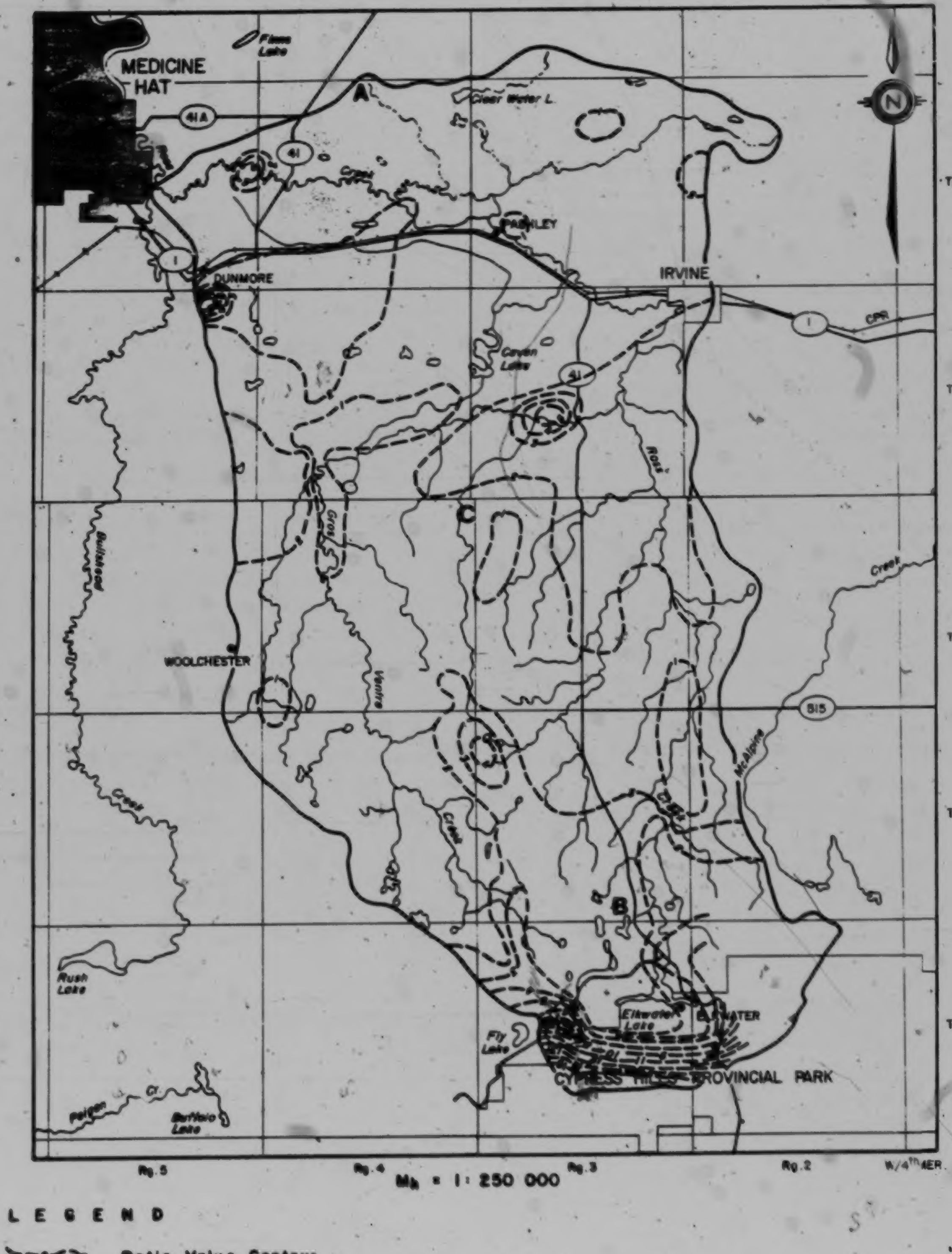
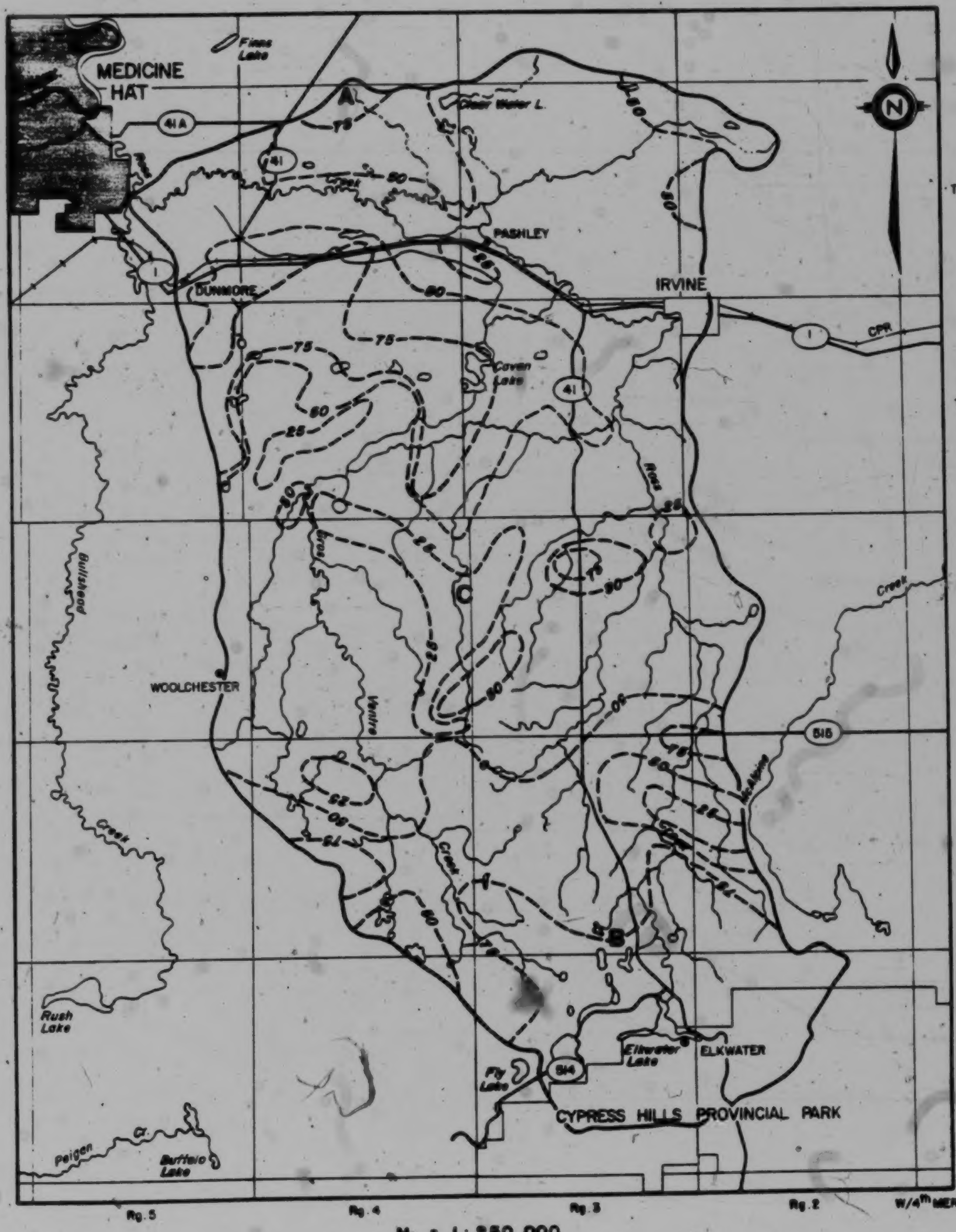

MEDICINE HAT
Flaxe Lake
A
Clear Water L.
41A
41
Creek
PARKLEY
1
DUNMORE
IRVINE
CPR
Cavan Lake
41
C
Ross
Gros
Bullshead
WOOLCHESTER
Ventre
Creek
515
McAlpine
Creek
B
Rush Lake
Fly Lake
Elkwater Lake
ELKWATER
CYPRESS HILLS PROVINCIAL PARK
Pelican Cr.
Buffalo Lake
N
Tp 12
Tp 11
Tp 10
Tp. 9
Tp. 8
Rg. 5
Rg. 4
Rg. 3
Rg. 2
W/4TH MER.
Mh = 1 : 250 000
LEGEND
Ratio Value Contour

Figure 18. Areal distribution of $Ca^{2+}$ to $Mg^{2+}$ ratio.


MEDICINE HAT
Finns Lake
Clear Water L.
41A
41
PASHLEY
IRVINE
DUNMORE
1
CPR
Coven Lake
Ross
WOOLCHESTER
Bullshead
Ventre
Creek
516
Mc Alpine
Rush Lake
Fly Lake
514
Elkwater Lake
ELKWATER
CYPRESS HILLS PROVINCIAL PARK
Peigan Cr.
Buffalo Lake
A
B
C
75
50
25
Tp 12
Tp 11
Tp 10
Tp 9
Tp 8
Rg. 5
Rg. 4
Rg. 3
Rg. 2
W/4th MER
N
$M_h$ = 1 : 250 000

L E G E N D

—25— Percent Bicarbonate & Carbonate Value Contour

Figure 19. Areal distribution of bicarbonate and carbonate.

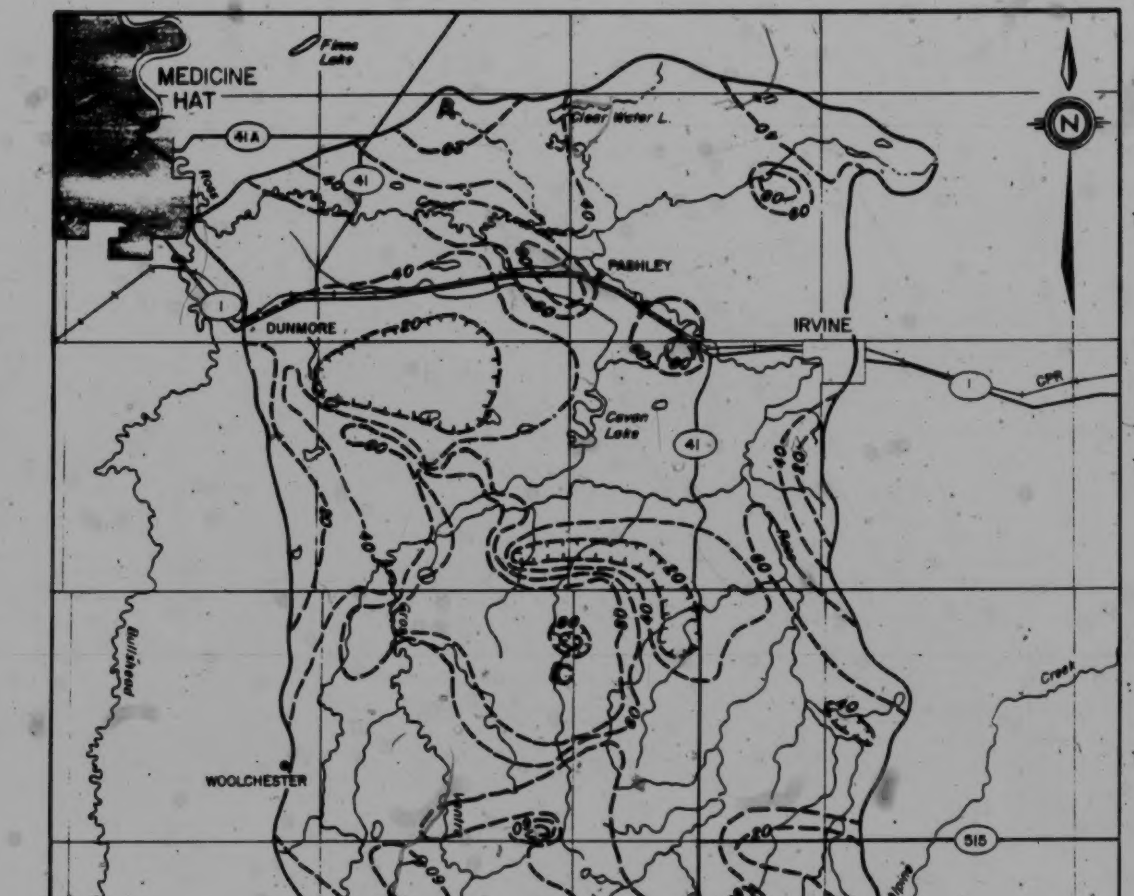

MEDICINE HAT
Flaxe Lake
Clear Water L.
41A
41
1
PASHLEY
DUNMORE
IRVINE
CPR
Cavan Lake
Bullshead
WOOLCHESTER
515
McAlpine
Creek
Rush Lake
Fly Lake
514
Elkwater Lake
ELKWATER
CYPRESS HILLS PROVINCIAL PARK
Pelican Cr.
Buffalo Lake
A
B
C
Tp.12
Tp.11
Tp.10
Tp.9
Tp.8
Rg.5
Rg.4
Rg.3
Rg.2
W/4th MER
20
40
60
80
N
Mh = 1 : 250 000
LEGEND
60 Sulphate Percent Value Contour

Figure 20. Areal distribution of sulphate.

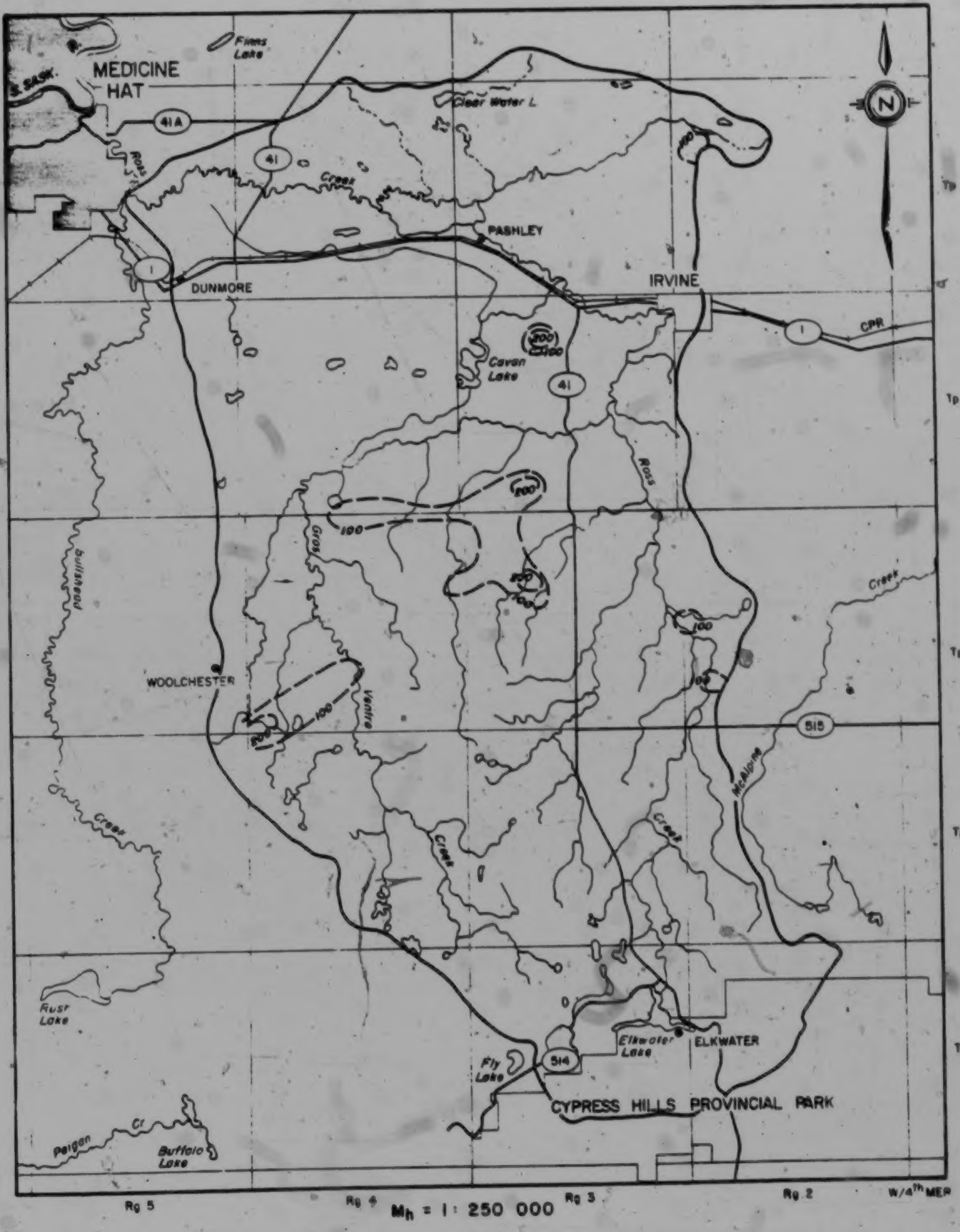

MEDICINE HAT
S. SASK.
Finns Lake
Clear Water L.
41A
41
Ross
Creek
PASHLEY
1
DUNMORE
IRVINE
CPR
200
100
Cavan Lake
Ross
Gros
Bullshead
WOOLCHESTER
Ventre
515
McAlpine
Creek
Rust Lake
Elkwater Lake
ELKWATER
Fly Lake
514
CYPRESS HILLS PROVINCIAL PARK
Paigan Cr
Buffalo Lake
Tp 12
Tp 11
Tp 10
Tp 9
Tp 8
Rg 5
Rg 4
Rg 3
Rg 2
W/4th MER
Mh = 1: 250 000
N
L E G E N D
—200— Chloride Value Contours

Figure 21. Areal distribution of chloride.

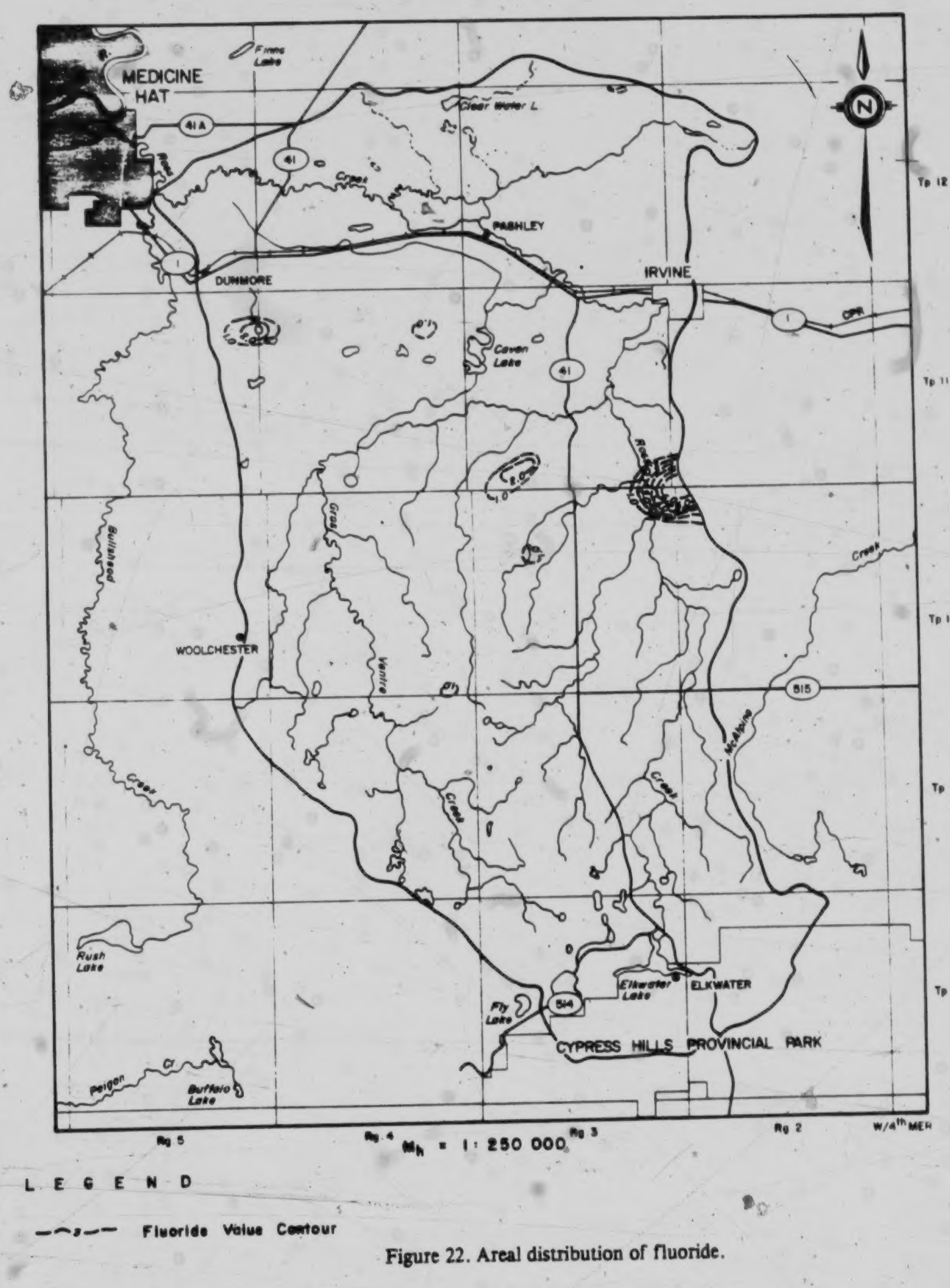
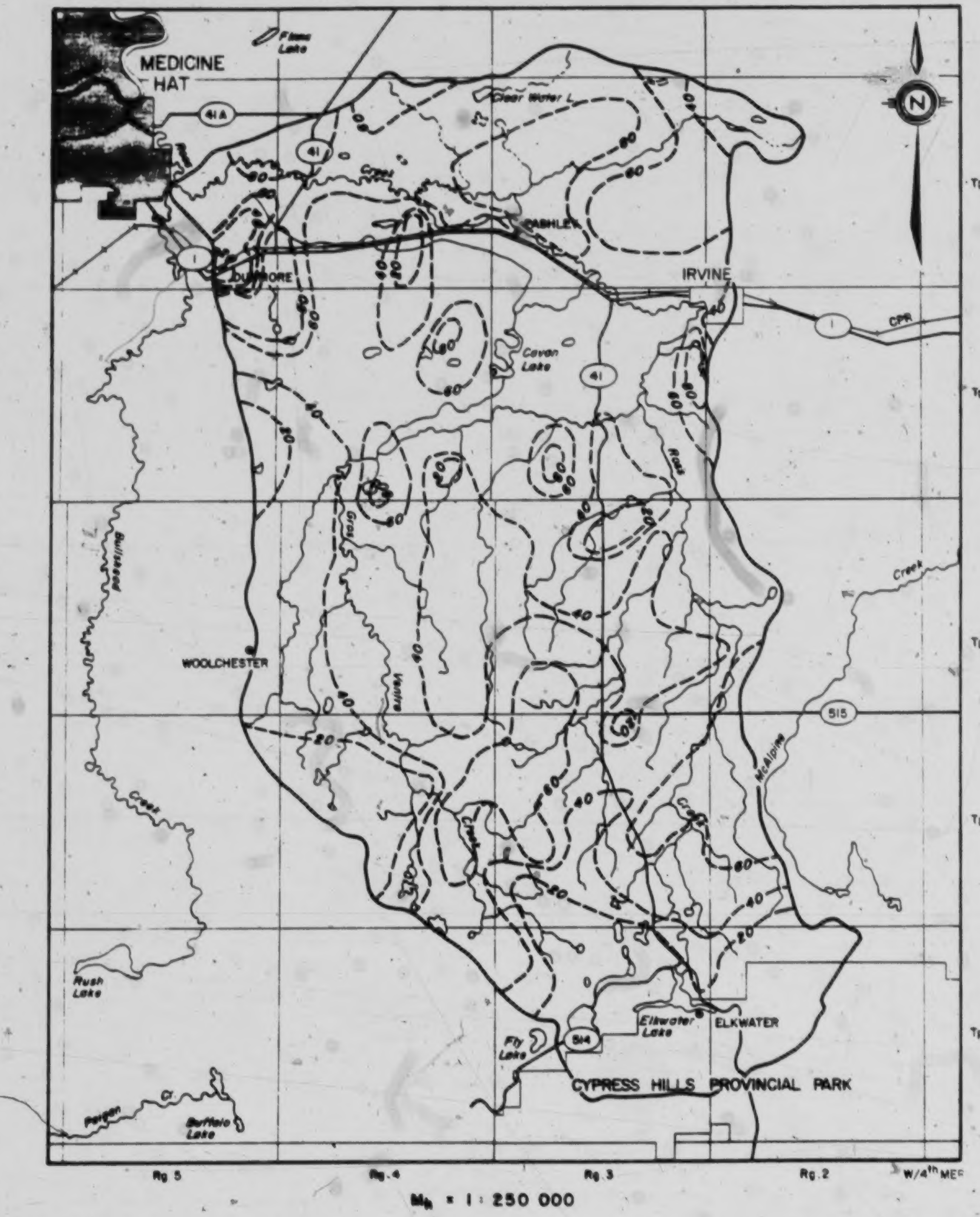

MEDICINE HAT
Finns Lake
Clear Water L.
41A
41
Creek
PASHLEY
IRVINE
1
DUNMORE
CPR
Coven Lake
Tp 12
Tp 11
Tp 10
Tp 9
Tp 8
Bullshead
Gros
Ventre
WOOLCHESTER
Creek
515
McAlpine
Creek
Creek
Rush Lake
Fly Lake
514
Elkwater Lake
ELKWATER
CYPRESS HILLS PROVINCIAL PARK
Peigan Cr
Buffalo Lake
Rg 5
Rg 4
Rg 3
Rg 2
W/4th MER
$M_h$ = 1: 250 000
1.0
2.0
L E G E N D
Fluoride Value Contour

Figure 22. Areal distribution of fluoride.


MEDICINE HAT
Flaxe Lake
41A
41
Clear Water L.
Creek
ASHLEY
DUNMORE
1
IRVINE
CPR
Cavan Lake
Ross
Bullshead
WOOLCHESTER
Gros
Ventre
Creek
515
McAlpine
Creek
Rush Lake
Fly Lake
514
Elkwater Lake
ELKWATER
CYPRESS HILLS PROVINCIAL PARK
Pelgan Cr.
Buffalo Lake
20
40
60
Tp 12
Tp 11
Tp 10
Tp 9
Tp 8
Rg 5
Rg 4
Rg 3
Rg 2
W/4th MER
Mh = 1 : 250 000

L E G E N D
— —40— — Percentage Sodium & Potassium Value Contour

Figure 23. Areal distribution of sodium and potassium.

distribution, except that: water north of the Trans-Canada Highway has an average content of 60 percent; and the content increases northwards from the Cypress Hills in the southern one-quarter of the map-sheet.

Generally, iron occurs in amounts less than 1 mg/l in the groundwater of the map-sheet (Figure 24), and values greater than 1 mg/l are seldom recorded. Iron occurrences greater than 12 mg/l were determined in two wells with corroding casings, and are not considered as part of the natural distribution of iron content in the water of the basin.

### 2.7.3 Chemical types or facies of groundwater

Two Piper-type trilinear diagrams (Piper, 1944) have been constructed from the hydrochemical analyses. For convenience, chemical data from wells only are presented in Figure 25, whereas spring and surface water data are shown in Figure 26. Since only a few wells are known to bottom in the bedrock strata, no distinction has been made regarding the variation in well depths in constructing the figures.

Based on the major anion and cation content of water in the map-area and arbitrarily selected boundary lines at the 40 percentile of $SO_4^{2-} + Cl^-$ and $Ca^{2+} + Mg^{2+}$ ppm, four hydrochemical facies have been identified as below (Figure 27):

1. $Ca$-$Mg$-$SO_4$-$HCO_3$ facies;
2. $Ca$-$Mg$-$HCO_3$ facies;
3. $Na$-$SO_4$-$HCO_3$ facies;
4. $Na$-$HCO_3$ facies.

The relationships between groundwater chemistry, and groundwater environment and flow patterns are better explained with the hydrochemical facies concept of Back (1960, 1966).

Most of the groundwater in Ross Creek Basin occurs as facies 1 ($Ca$-$Mg$-$SO_4$-$HCO_3$), which has an average total dissolved solids value of 2 028 ppm, and consists typically of $Ca^{2+} + Mg^{2+}$ greater than 40 percent total cations. Sulphate is usually the major anion, but bicarbonate and carbonate content ranges between 8 and 58 percent of the total anions in waters of this facies. This type of water tends to occur mainly along Ross and Gros Ventre Creeks, and their major tributary channels.

Facies 2 ($Ca$-$Mg$-$HCO_3$) waters are the next most abundant (Figure 25), although they appear to be the most abundant in the basin according to Figure 26. Having a low average total

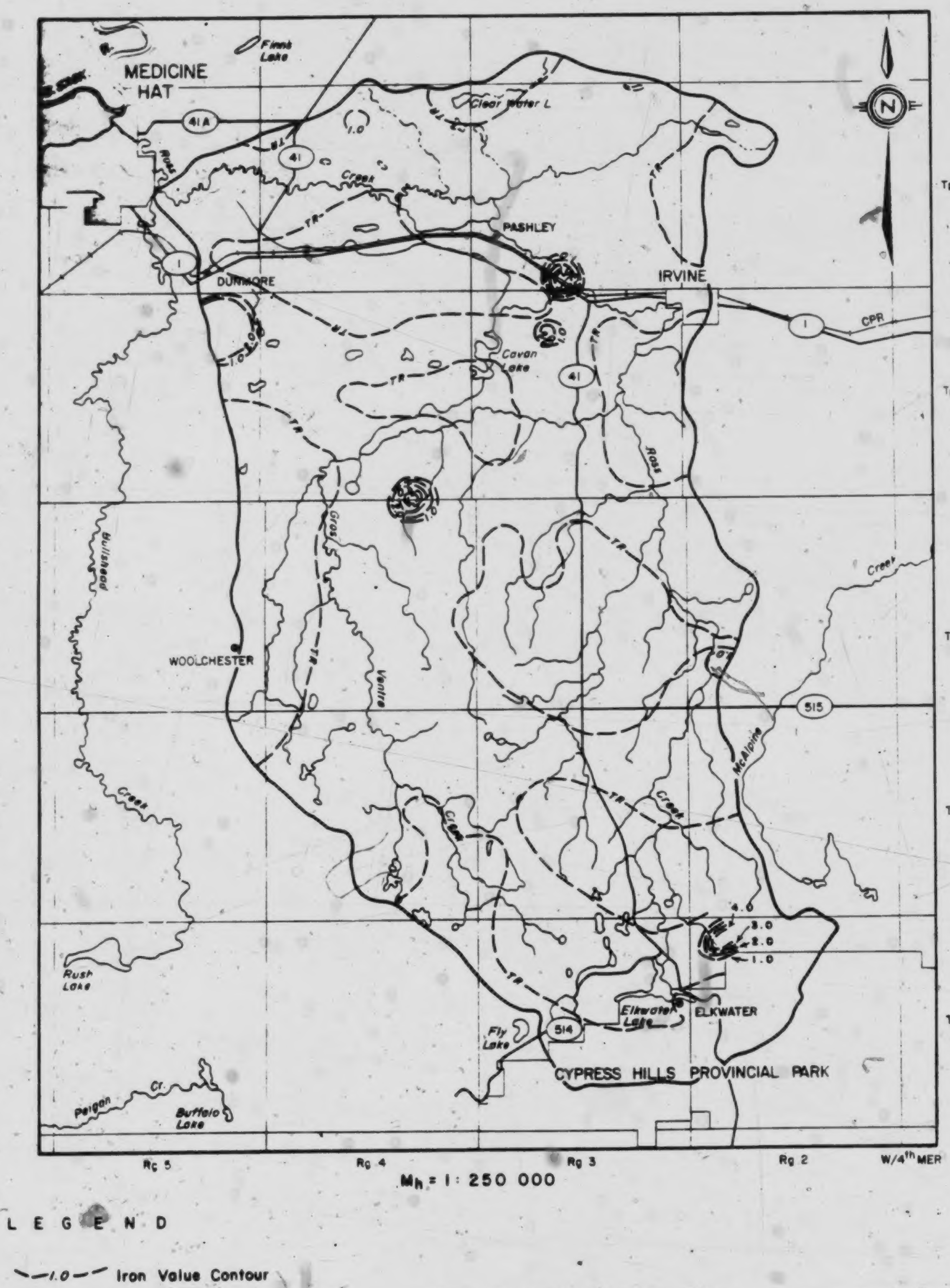
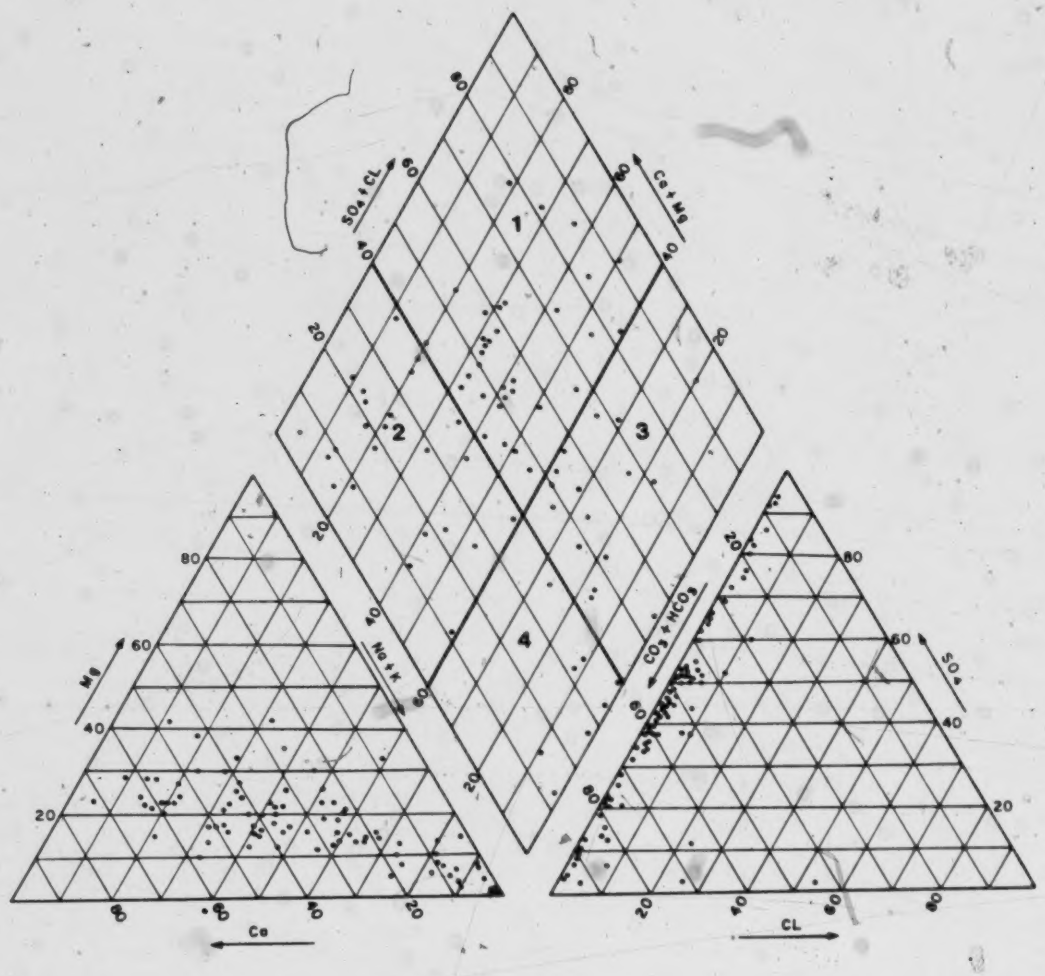

Finns Lake
MEDICINE HAT
Clear Water L
41A
41
1.0
Ross
Creek
TR
PASHLEY
IRVINE
DUNMORE
1
CPR
Cavan Lake
41
Ross
Bullshead
Gros
Ventre
WOOLCHESTER
Creek
515
McAlpine
Creek
4.0
3.0
2.0
1.0
Rush Lake
Elkwater Lake
ELKWATER
Fly Lake
514
CYPRESS HILLS PROVINCIAL PARK
Peigan Cr.
Buffalo Lake
Tp 12
Tp 11
Tp 10
Tp 9
Tp 8
Rg 5
Rg 4
Rg 3
Rg 2
W/4th MER
N
Mh = 1: 250 000
LEGEND
1.0 Iron Value Contour

Figure 24. Areal distribution of iron.



PERCENT OF TOTAL PARTS PER MILLION

Figure 25. Trilinear diagram showing chemical compositions of well water samples.

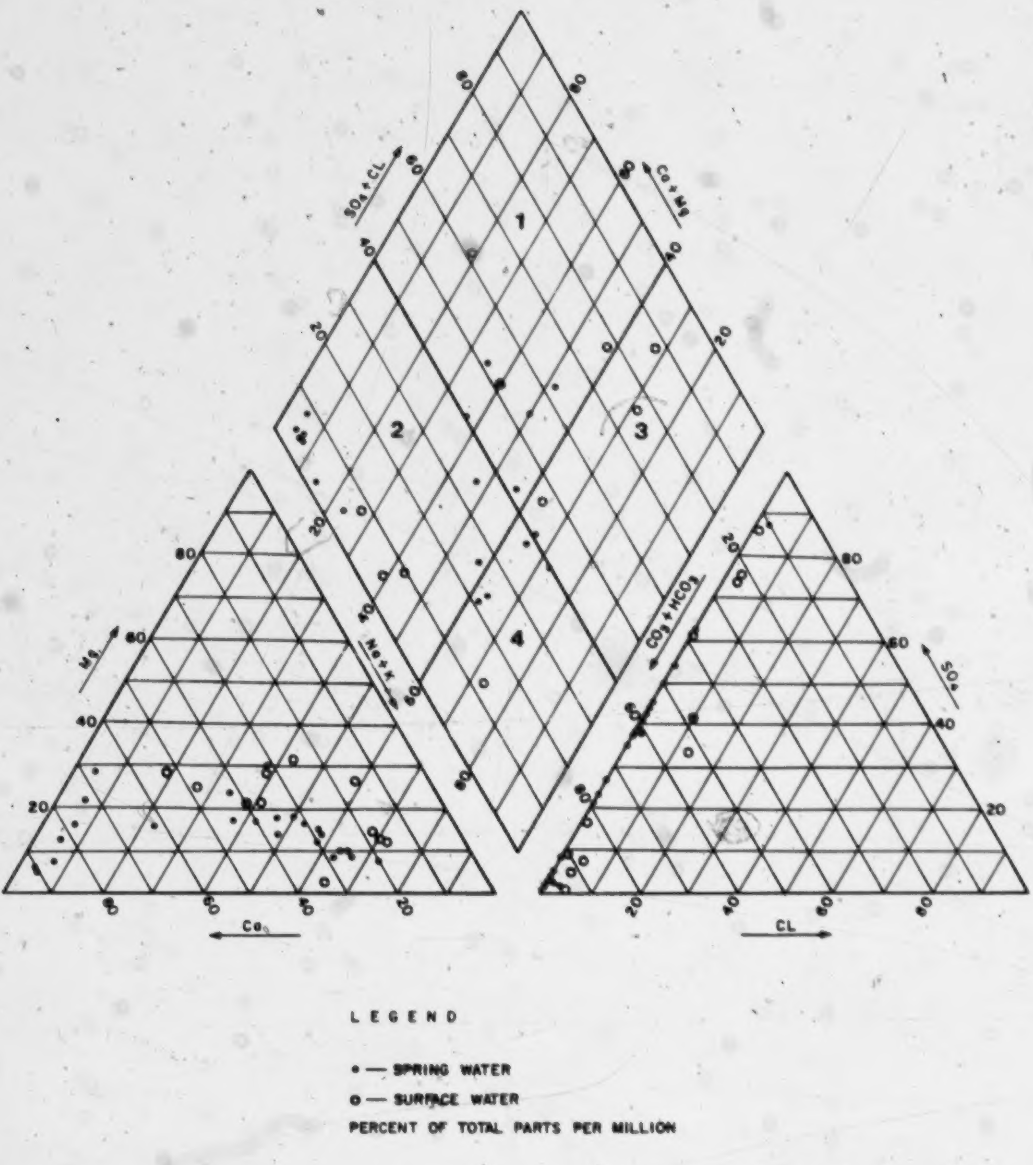


Figure 26. Trilinear diagram showing chemical compositions of spring and surface water samples

dissolved solids value of 466 ppm. $Ca^{2+}$ + $Mg^{2+}$ are the major cations of this facies, whereas $HCO_3^-$ constitutes at least 60 percent of the total anion content. These waters dominate the area north of the Cypress Hills, the area south of the Trans-Canada Highway and west of Cavan Lake, and occur as discontinuous patches along the western and northern drainage divides.

Facies 3 ($Na$-$SO_4$-$HCO_3$) waters are next to facies 2 waters in abundance (Figure 25). These waters are characterized by a high mean total dissolved solids value of 3 317 ppm, and $Na^+$ occurs in amounts greater than 60 percent of the total cation content. Sulphate and bicarbonate which constitute more than 40 percent and less than 60 percent of the total anions respectively, are the main anions of facies 3 waters. Waters in this group occur in isolated zones all over the basin.

The least encountered type of water in the basin belongs to facies 4 ($Na$-$HCO_3$). The major cation and anion are sodium and bicarbonate which together with minor magnesium and calcium, and sulphate ions yield an average total dissolved solids value of 972 ppm in these waters. Like type 3 waters, water belonging to this facies is distributed sporadically across the basin, but it usually occurs in close association with type 2 waters.

### 2.7.4 Interpretation of hydrogeochemical results

A geochemical analysis of groundwater is important not only for determining the potability of groundwater, but also in understanding the geology and movement of groundwater. In this section, an attempt is made to relate the geochemical results in the preceding paragraphs to the field-mapped groundwater flow distribution (Figure 12), and to the suitability of groundwater in the map-area for human consumption.

Various workers have demonstrated the relationship between groundwater chemistry, and the hydrogeological environment and groundwater flow. Chebotarev (1955) proposed regional sequential changes in dominant anions along flow lines as:

$$HCO_3^- \rightarrow HCO_3^- + SO_4^{2-} \rightarrow SO_4^{2-} + HCO_3^- \rightarrow SO_4^{2-} + Cl^- \rightarrow$$
$$Cl^- + SO_4^{2-} \rightarrow Cl^-$$

Tóth (1966b) indicated an apparent tendency for waters in a flow system to shift their chemical character from a pure bicarbonate type toward the sulphate facies whilst moving to the areas of upward flow. In large sedimentary basins, the Cheboratev sequence can be correlated with depth (Domenico, 1972), and have been described in terms of three main zones as (Freeze and Cherry, 1979): upper zone of mainly $HCO_3^-$ water, low in total dissolved solids; intermediate zone of higher total dissolved solids and $SO_4^{2-}$ as the major anion; and lower zone with high total dissolved solids and high concentration of $Cl^-$. All these studies lead to a common conclusion that: with allowances for interruption and incompletion resulting from the scale and geology of a specific setting, groundwater should evolve naturally from a bicarbonate facies in recharge areas to sulphate- or chloride-dominated facies in discharge areas.

It is now possible to attempt a correlation of the various chemical facies (Figure 27), and therefore individual chemical constituents with areas of recharge and discharge in the mapped flow systems (Figure 12). In general, the hydrochemical facies distribution in the area of study appear to reflect the envisaged groundwater flow patterns.

A comparison of Figures 12 and 27 shows that facies 1 ($Ca$-$Mg$-$SO_4$-$HCO_3$) and facies 3 ($Na$-$SO_4$-$HCO_3$) groundwaters are dominant in the apparent discharge areas, whereas the "young waters" of facies 2 ($Ca$-$Mg$-$HCO_3$) and 4 ($Na$-$HCO_3$) occur mainly in the mapped recharge and midline areas. This situation is in agreement with the propositions of Chebotarev (1955), Tóth (1966b), Domenico (1972), and Freeze and Cherry (1979). Bicarbonate, the major anion component of waters in recharge areas is formed through the following process: meteoric water ($H_2O$) dissolves carbon dioxide ($CO_2$) in the atmosphere on its way to the lithosphere to produce carbononic acid ($H_2CO_3$), which is enhanced in the $CO_2$-rich root zone by further dissolution of $CO_2$ generated by the decay of organic matter, and by the respiration of plant roots. Though weak, carbonic acid is capable of dissolving calcium carbonate ($CaCO_3$) along flow paths of the infiltrating water. This sequence of events results in excess $HCO_3^-$ in recharge areas as (Davis and De Wiest, 1966):

$$H_2O + CO_2 \rightarrow H_2CO_3 \rightarrow H^+ + HCO_3^-$$

$$HCO_3^- \rightarrow H^+ + CO_3^{2-}$$

$$H^+ + CaCO_3 \rightarrow HCO_3^- + Ca^{2+}$$

As the water moves progressively towards the discharge areas, the rate of increase of $CO_3^{2-} + HCO_3^-$ decreases (Tóth, 1966b) at the expense of $SO_4^{2-}$, which is acquired readily from the bedrock strata. The high mean total dissolved solids values of facies 1 and 3 waters

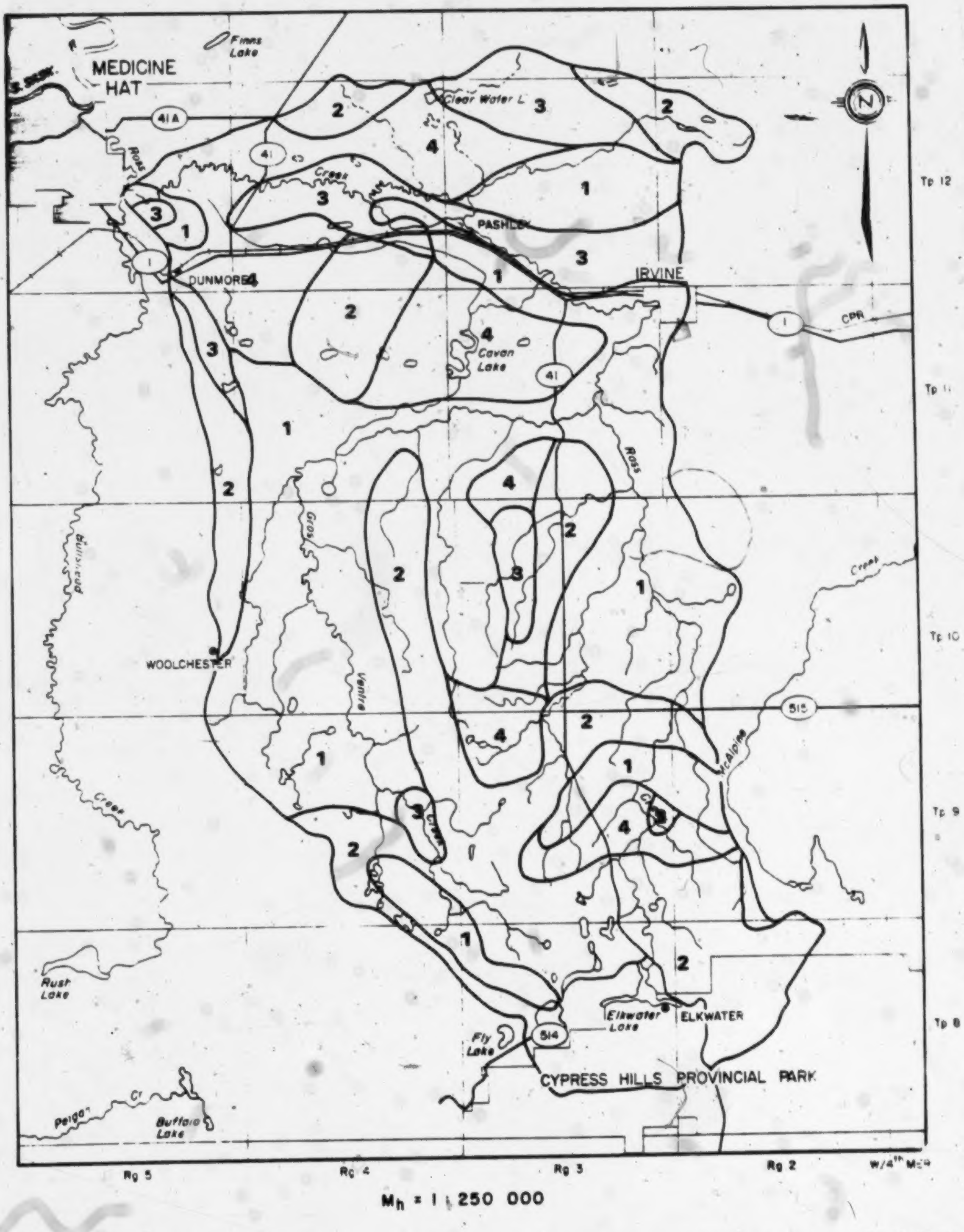
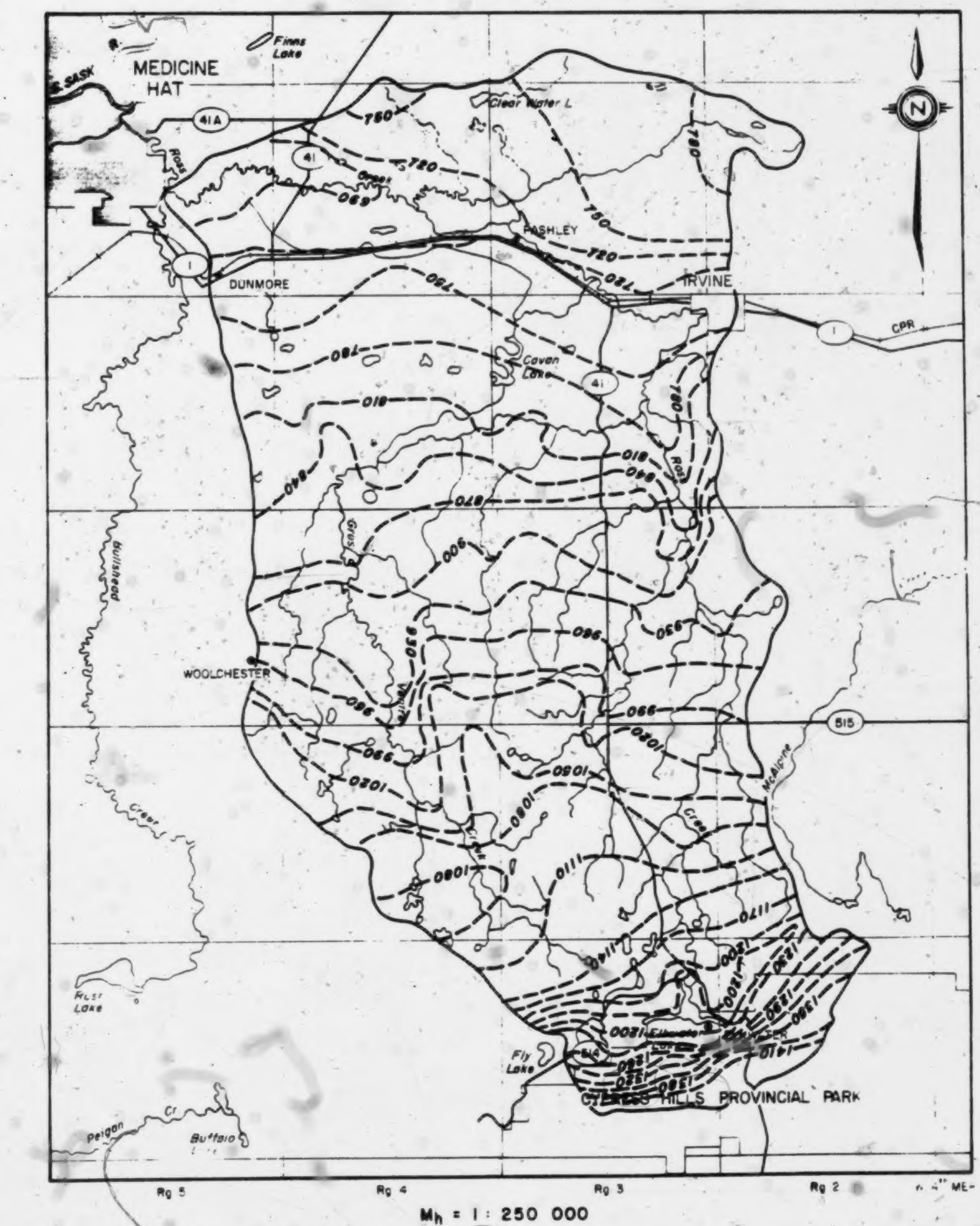

MEDICINE HAT
S. Sask.
Finns Lake
Clear Water L
41A
41
1
515
514
Ross
Creek
PASHLEY
IRVINE
DUNMORE
CPR
Cavan Lake
Ross
Creek
Bullshead
Gros Ventre
WOOLCHESTER
McAlpine
Rush Lake
Fly Lake
Elkwater Lake
ELKWATER
CYPRESS HILLS PROVINCIAL PARK
Peigan Cr
Buffalo Lake
N
Tp 12
Tp 11
Tp 10
Tp 9
Tp 8
Rg 5
Rg 4
Rg 3
Rg 2
W/4th MER
Mh = 1 : 250 000
LEGEND
Facies Boundaries
1 Calcium, magnesium, sulphate, biocarbonate
2 — Calcium, magnesium, biocarbonate
3 — Sodium, sulphate, biocarbonate
4 — Sodium, biocarbonate

Figure 27. Areal distribution of hydrochemical facies in Ross Creek Basin.

suggest an addition of new mineral matter to the groundwater, and hence are indicative of upward moving groundwaters where these facies occur.

A careful observation of Figures 17, 18, 19 and 20 reveals some noteworthy points: 1. areas of low total dissolved solids appear to coincide with those of high $Ca^{2+}$ : $Mg^{2+}$ ratio, low percentage of $SO_4^{2-}$, and high percentage of $HCO_3^-$ (e.g. points A and B); these areas occupy the highest topographic elevations in their immediate environs, and are associated with recharge areas in Figure 12; 2. areas of high total dissolved solids often coincide with those of low $Ca^{2+}$ : $Mg^{2+}$ ratio, high percentage of $SO_4^{2-}$ and low percentage of $CO_3^{2-}$ + $HCO_3^-$ (e.g. point C), and such areas occur along low topographic elevations in discharge areas (Figure 12); 3. in general, the direction of gradual increase in total dissolved solids tends to correspond with that of increasing percent $SO_4^{2-}$, decreasing percent $CO_3^{2-}$ + $HCO_3^-$ and $Ca^{2+}$ : $Mg^{2+}$ ratio, and thus may indicate direction of groundwater movement in groundwater flow systems; 4. mineralization of groundwater attains its maximum value in the area around point C (Figure 17). This is probably due to the retarding effect or even stagnation of groundwater caused by downward-moving water down slope of point C and around Cavan Lake, and assisted by westward-moving groundwater along the lower Ross Creek valley (Figure 28); 5. along the Ross Creek valley, mineralization appears to be lower than expected for its postulated position as a regional discharge area; this situation can result from either of rapid flushing of discharged high total solids water away from the basin by fast moving groundwater in the considerably permeable rock units below and along the Ross Creek valley, the neutralizing effect of "fresh" recharging water from the adjacent local flow systems on both sides of Ross Creek, or the stagnation of the high total dissolved solids water around point C, which will otherwise be discharged along Ross Creek valley.

The simple relationship between hydrochemical constituents and flow patterns as portrayed by the first three points has been found to be more complicated in several regional groundwater studies. Since these complications are only manifested principally by imperfect coincidence of boundary lines of the flow systems and those of hydrochemical facies and constituents in Ross Creek Basin, their detailed discussion is not included in this study. In any event, such complications may be caused by variation in scale and chemical composition of the host rock; geochemical processes, such as cation exchange, carbon dioxide generation in the saturated zone, and oxidation-reduction of sulphates and sulphide minerals; and the highly


Finns Lake
MEDICINE HAT
S. SASK
41A
41
1
515
Clear Water L
Ross
Creek
PASHLEY
DUNMORE
IRVINE
CPR
Cavan Lake
Bullshead
Gros Ventre
WOOLCHESTER
McAlpine
Rost Lake
Fly Lake
Elkwater Lake
ELKWATER
CYPRESS HILLS PROVINCIAL PARK
Peigan Cr
Buffalo Lake
750
720
690
780
810
840
870
900
930
960
990
1020
1050
1080
1110
1170
1200
1230
1260
1290
1320
1350
1380
1410
Tp 10
Tp 9
Tp 8
Rg 5
Rg 4
Rg 3
Rg 2
W. 4th MER.
N
Mh = 1 : 250 000

L E G E N D
—900— Contour Line
( Surface Contour Interval 30 metres –
Elevation in metres Above Sea Level )

Figure 28. Water table map.

variable scale, complexity, and the often time-dependent nature of the prairie groundwater flow systems as discussed in Freeze and Cherry (1979), and Garven (1980).

In general, evidence from the hydrogeochemical analyses of groundwater samples in the map-area is in good agreement with the picture of groundwater movement (Figure 12) obtained as the final product of the hydrogeological mapping.

The potability of water in Ross Creek Basin may be comprehended easily from Figures 17 to 24, and Table 7 which shows the Alberta Public Health Units standards for potable water. Most of the groundwater in the basin is suitable for human consumption, at least, within treatable limits. In a few areas, however, where the total dissolved solids are exceedingly high, it might be possible to consider water for other uses, such as stockwatering, irrigation, industry, and so on. It is noteworthy that, very often, chemical standards for potable water are controlled by the quality of available water.

### 2.8 Dynamic reserves of groundwater

The dynamic reserves of groundwater constitute that portion of water which recharges and moves through the groundwater basin, and which may be withdrawn without causing a permanent drop in the water levels. In effect, it is the safe yield of the basin, and it corresponds to the natural basin yield defined by Freeze (1969b) as "the quantity of flow through an undeveloped basin under natural conditions".

Several methods are employed in the estimation of regional groundwater recharge, including the baseflow method, recharge method, budget method, theoretical calculation of basin yield, and the dynamic reserves method. The dynamic reserves method used herein employs the isolated exchange of groundwater between areas of downward flow and areas of upward flow. This condition is approximated closely during periods of ground frost, when recharge by precipitation and discharge by evapotranspiration are negligible (Tóth, 1968, 1982).

The natural yield Q(i,t) of any flow system i, during a given time interval t, equals the amount of water passing through and at right angles to its cross-section at the boundary between the area of downward flow and the area of upward flow. For a flow system of equal areas of upward and downward flow, the natural yield for the time interval is expressed as (Tóth, 1968):

Table 7. Chemical quality standards of the Alberta Public Health unit.

| Constituents | Concentration |
|---|---|
| Total solids | 1 000 - 2 000 mg/l |
| Calcium | 200 mg/l |
| Magnesium | 150 mg/l |
| Sodium | 300 mg/l |
| Iron | 3 mg/l |
| Sulphates | 1 000 mg/l |
| Chlorides | 300 - 500 mg/l |
| Fluoride | 1.5 mg/l |
| Alkalinity | 30 - 50 mg/l |
| Hardness | 400 (plus very hard) |
| NO | 0.1 mg/l |
| Nitrates | 10 mg/l |
| pH | 7 (alkaline and acidic - high acidity, high alkalinity) |

$$Qit = \bar{S}y . \bar{f}_t \frac{Ai}{2} \qquad (4)$$

where: $Ai$ = total land area of the flow system [$L^2$];

$Qit$ = natural yield [$L^3$];

$\bar{f}_t$ = average change of water level in either the downward or upward area [L];

$\bar{S}y$ = specific yield [dimensionless].

Water levels were measured at 18 observation points in the basin for a period of approximately one year, including the normal five-month period of ground frost (November to March) in the Plains Regions of Alberta. The details of the water measurements are reported in Ophori and Tóth (1985). From the weekly and biweekly measurements in the observation wells of both recharge and discharge areas, the average change in water levels for the five-month period of ground frost was found to be (Figure 29):

$$\bar{f}_{t = 5 \text{ months}} = 0.47 \text{ m}$$

If it is assumed that the specific yield of the local rocks is 0.1, then the average depth of water exchange for a period of 5 months is:

$$q_{t=5} = \bar{S}y \times \bar{f}_{t=5} = 0.1 \times 0.47 = 0.047/5 \text{ months}$$

Thus, the natural yield of the flow system per unit area per year ($q_T$) may be calculated as:

$$q_T = 0.047 \times \frac{12}{5} = 0.113 \text{ m/year}$$

The groundwater coefficient of precipitation (Kp) is (Tóth, 1968):

$$Kp = \frac{q_T}{p}$$

where: p = the amount of annual precipitation.

Recalling that the annual precipitation in the basin is 531 mm (section 2.3.1), the groundwater coefficient of precipitation will then be:

$$Kp = \frac{113}{531} \times 100 = 21 \text{ percent}$$

Inherent in the calculations is the assumption that the sizes of the areas of downward flow and the areas of upward flow are equal. This assumption is fairly satisfied by the earlier calculation of these sizes (section 2.4). Thus 21 percent of the annual precipitation is the quantity of water moving annually from recharge areas to discharge areas in Ross Creek Basin.

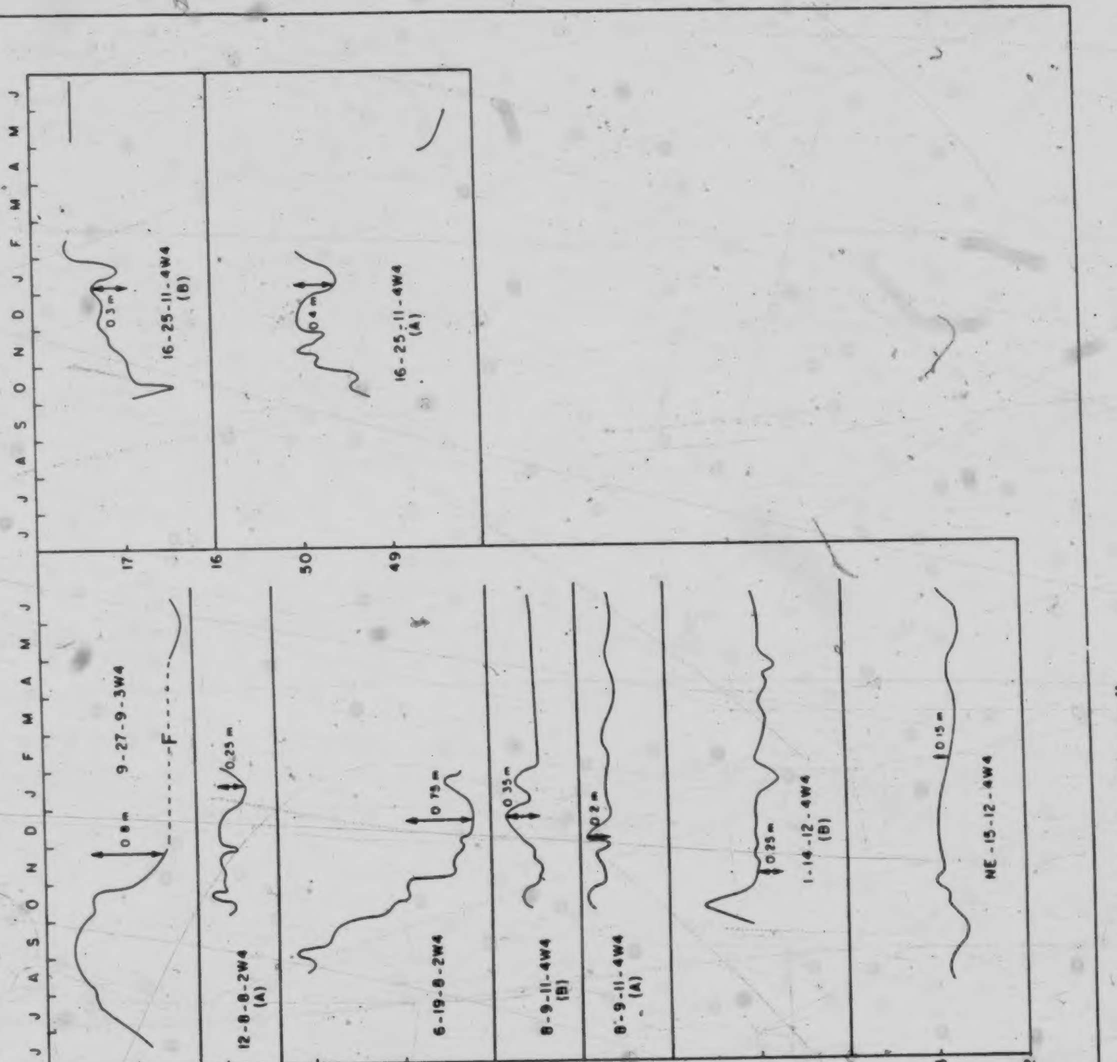


Figure 29. Water levels in observation wells.

Furthermore, it means that an average depth of water of 113 mm per year or $3.6x10^{-3}$ $m^3/s/km^2$ recharges the zone of saturated flow in the recharge areas during the time intervals specified. The calculated groundwater coefficient of precipitation (21%) is close to average values (2-20%) estimated in other parts of Alberta (Farvolden, 1963; Tóth, 1968, 1982; Akiti, 1975).

Often, withdrawal is distributed over both areas of upward and downward flow, thus, the evenly distributed safe production rate is half of the groundwater reserves or natural basin yield. In Ross Creek Basin, the safe production rate is $1.8x10^{-3}$ $m^3/s/km^2$. It must be remarked, however, that this method of estimation of production rates does not account for possible variations in geology across the basin.

### 2.9 Surface water hydrology

Since surface water is an integral part of the hydrologic cycle, it is also important in a basin-wide evaluation of subsurface hydrology. Although the availability of long-term quantitative information on surface water in Ross Creek Basin is limited, several hydrological reports are available to provide background information for the present study.

There are two hydrometric gauging stations with sufficiently long quantitative streamflow records in the study area. The first station was established on Ross Creek (05AH003) near Irvine (NW-34-11-3-4) where flows were recorded during 1910-1931 and 1935-1985, whereas the second station located on Gros Ventre Creek (05AH037) near Dunmore (SE-8-11-4-4) has flow measurements from 1921-1922 and 1944-1983. These records have been analyzed extensively by various workers through time who reported annual runoff values of $9.7x10^6$ $m^3$ (PFRA, 1956), $13.9x10^6$ $m^3$ (Ashraf, 1978), $13.1x10^6$ $m^3$ (Keith Consulting, 1979), and 9.3 to $14.2x10^6$ $m^3$ (Northwest Hydraulic Consultant Ltd., 1981a). These values convert to an average annual runoff of 20 mm or 3.8 percent of the average annual precipitation. Assuming that no groundwater moves out of the drainage basin, that is, the surface and groundwater boundaries are coincident, 96.2 percent of the precipitation will be available for evapotranspiration. Of this amount, 21 percent completes the subsurface portion of the hydrologic cycle, whereas 75.2 percent must evaporate or be transpired without participating in the groundwater circulation. Furthermore, a baseflow of essentially zero was estimated from streamflow records in the basin (Ophori, 1986), thus most of the 21 percent subsurface water is discharged and evapotranspired through the numerous springs.

Although 70 percent of the streamflow in the basin results from spring runoff during the months of march and April, the flow has an unpredictable nature (Keith Consulting, 1979). In general, the magnitude and date of floods have varied from year to year, and spring runoff was insignificant for several years.

Cavan and Elkwater Lakes are the two largest surface water bodies in the basin. In a few cases, the storage capacities of these bodies have been analyzed by some of the studies referenced earlier. As a result of the hummocky character of the surficial deposits in the proximity of the Cypress Hills slopes, sloughs are abundant in this part of the basin. Sloughs affect the surface water hydrology in an area by limiting surface runoff and acting as very small closed drainage basins. They provide a significant amount of depressional storage whereby ponded water either infiltrates into the ground, transpired by hydrophytes and phreatophytes, or evaporates. Sloughs often complicate the simple relationship between surface water hydrology and the local groundwater regime (Tóth, 1966a; Meyboom 1966, 1967; Lissey, 1968; Freeze, 1969a, Sloan, 1972). The influence of sloughs on the groundwater regime in Ross Creek Basin is not known, but their possible influence is considered briefly in a later section.

### 2.10 Water use

If there is no use for water, it will be valueless, thus, knowing the use for water is as important as exploring for it. The first water usage in the basin was evaluated by Keith Consulting (1979) as $3.36x10^{-1}$ m$^3$/s for the year 1977. In this evaluation, licensed or designed withdrawals rather than actual usage was considered. To gain an idea of actual water usage in the basin between 1978 and 1980, Alberta Environment (1980), by means of questionnaires, estimated actual withdrawals over the 3-year period of $3.5x10^{-1}$ m$^3$/s.

In the present study, a knowledge of the demand for water rather than actual usage is preferred, because it is on this basis that the role of available groundwater in the basin can be assessed. For this purpose, total authorized withdrawals by active licenses and pending applications were estimated from the 1984 files of the Water Rights Branch of Alberta Environment. A total water demand of $3.3x10^{-1}$ m$^3$/s was calculated for the basin through this process. Of this amount, $2.6x10^{-1}$ m$^3$/s was required for irrigation, $5.8x10^{-2}$ m$^3$/s for domestic use, and $6.8x10^{-3}$ m$^3$/s for municipal purpose.

# 3 GROUNDWATER MANAGEMENT

## 3.1 Aspects of management

Strictly speaking, the discussion of the aspects of groundwater management is a summary of the findings about the hydrogeologic environment, the groundwater regime, and the general nature of water in the basin from the literature and field stage.

Cross sections across the basin (Figure 2) reveal a topography whose slope is dominated in one direction (north-south direction). The east-west component of groundwater flow is, therefore, assumed to be negligible regionally compared to the north-south component; that is, a two-dimensional cross-sectional flow is assumed. Although the basin consists of different formations geologically, an analysis based on hydrogeologic units defines two different, but hydraulically uniform units, two aquifers and an aquitard. Whereas one of the aquifers is restricted to a little zone around the lower Ross Creek valley, the second aquifer lies in the property of the Cypress Hills Provincial Park. An analysis of water level fluctuation in observation wells produced a recharge from precipitation to the groundwater system of 113 mm or $3.0\times10^{-4}$ m/day. On the basis of topography, hydrogeologic units, and climate, a hydrogeologic environment is defined for Ross Creek Basin as illustrated in Figure 30. In general, Figure 30 lies between the ATR and BTR categories of Plains Regions environments (Ophori, 1986).

Mapping of surficial manifestations of groundwater movement reveals a groundwater regime of various orders of flow systems (local, intermediate and regional). Whereas springs and other discharge features mark the discharge ends of the flow systems, and the conspicuous absence of these features the recharge areas, a relatively large area has been mapped as uncertain on the basis of the features. Aquifer hydraulics shows the lack of reliable aquifers in the basin, and indicates that less than $4.1\times10^{-3}$ $m^3/s$ of groundwater can be produced from a few localities only, to cause complete mining of the resource in twenty years. A safe production rate of $1.8\times10^{-3}$ $m^3/s/km^2$ that would cause no permanent water level lowering was calculated from the analysis of dynamic reserves. In addition to the good agreement between the groundwater chemistry and the identified flow systems, up to 50 percent of the groundwater is potable, at least within treatable limits.

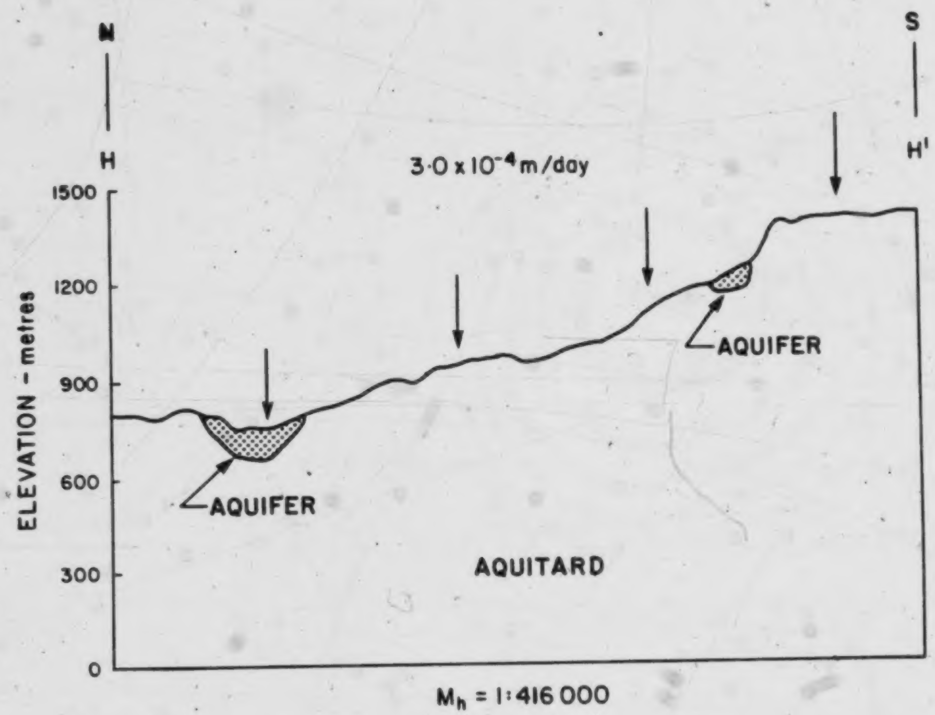

N
H
S
H'
3·0 x 10⁻⁴ m/day
1500
1200
900
600
300
0
ELEVATION - metres
AQUIFER
AQUIFER
AQUITARD
Mh = 1:416 000

Figure 30. [illegible]drogeologic environment of Ross Creek Basin.

Several issues call for groundwater management in Ross Creek Basin: 1. groundwater is scarce not because there is no recharge from precipitation, but because there is a lack in appropriate storage and transmission materials (aquifers) from which it can be produced as required; this is evident from the two very different groundwater productivity scenarios predicted on the basis of aquifer hydraulics on one hand, and dynamic reserves calculations on the other; 2. as a consequence of the first point above, several springs are discharging, evaporating and wasting groundwater, and there is the fear to issue licenses for their use because the groundwater regime is not understood; 3. runoff along Ross Creek and its tributaries is substantial relative to demands for water in the basin, but it is flashy in nature and requires storage for the time of need; and 4. annual, seasonal, and daily fluctuations in hydrometeorological studies of the basin are all in favour of a water management plan.

Under these conditions, the best use that can be made of the subsurface resources is their conjunctive operation with surface water resources. However, a prerequisite to such an operation is the presence of adequate aquifers that are capable of being recharged artificially or naturally through induced infiltration. The lack of such aquifers, according to our present knowledge of the basin, precludes a consideration of a conjunctive operation. Nevertheless, in order to make the best use of the available groundwater resources in the basin, a logical basis for the operation of the springs in the basin is formulated. The formulation is based on an understanding of the groundwater regime and its response to possible development.

### 3.2 Theoretical considerations

#### 3.2.1 Steady state groundwater regime

To gain a detailed understanding of the groundwater movement through the basin than permitted by field studies, the construction of a flow systems map was necessary. Unfortunately, the construction of meaningful hydraulic cross sections from field measurements of water levels was impossible because of the lack of observation points that are deep enough for this purpose. However, a theoretical flow pattern was determined for the basin with the aid of field observed data.

Figure 31 shows the mathematical model representation of Ross Creek Basin in two dimensions. As discussed previously, the section whose trace is indicated in Figure 1 was chosen

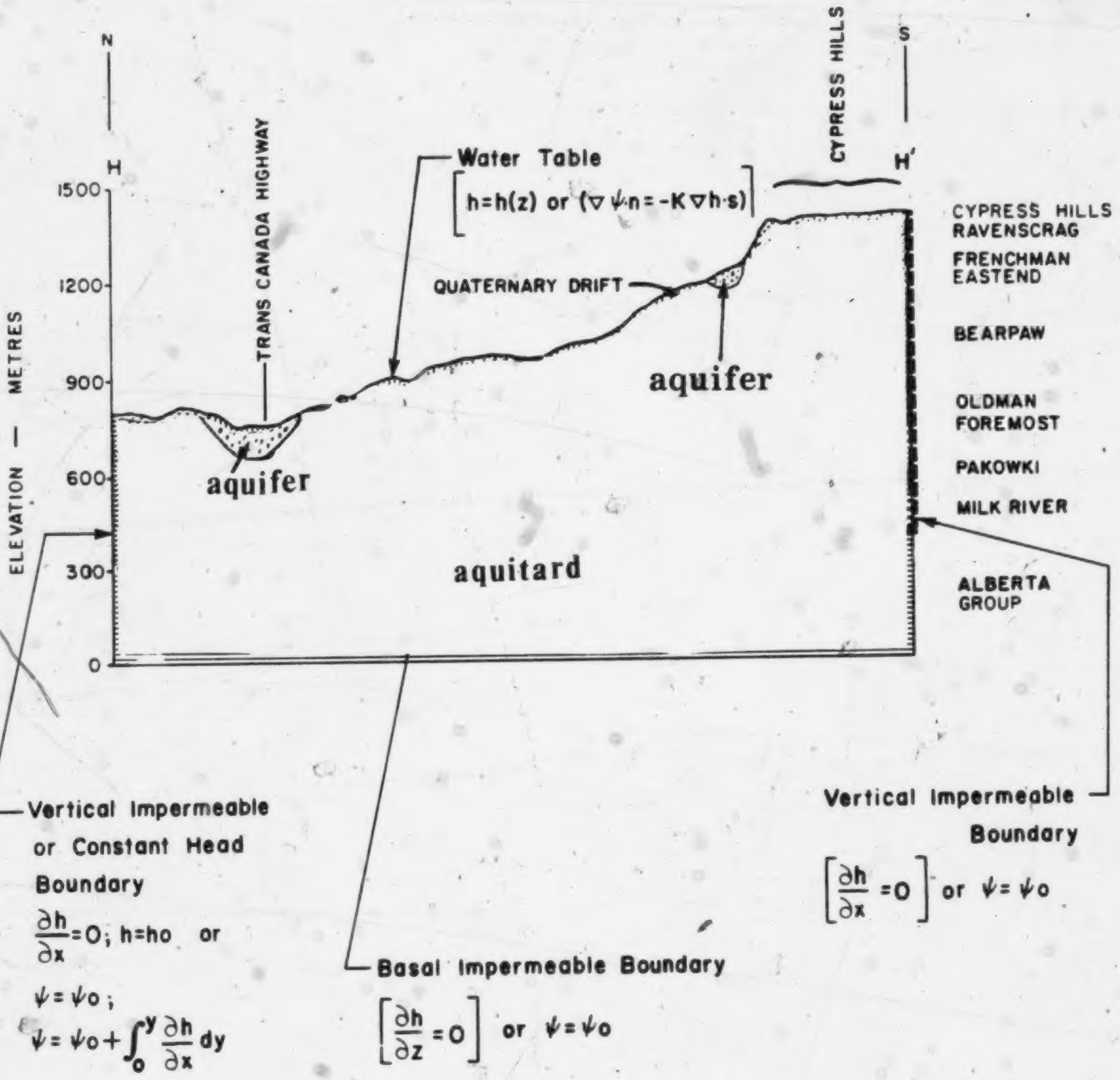

N
S
H
H′
CYPRESS HILLS
TRANS CANADA HIGHWAY
Water Table
[h=h(z) or (∇ψ·n=-K∇h·s)]
QUATERNARY DRIFT
aquifer
aquifer
aquitard
ELEVATION — METRES
1500
1200
900
600
300
0
CYPRESS HILLS
RAVENSCRAG
FRENCHMAN
EASTEND
BEARPAW
OLDMAN
FOREMOST
PAKOWKI
MILK RIVER
ALBERTA GROUP
Vertical Impermeable or Constant Head Boundary
$\frac{\partial h}{\partial x}=0$; h=ho or
$\psi=\psi_o$;
$\psi=\psi_o+\int_0^y \frac{\partial h}{\partial x}\,dy$
Basal Impermeable Boundary
$\left[\frac{\partial h}{\partial z}=0\right]$ or $\psi=\psi_o$
Vertical Impermeable Boundary
$\left[\frac{\partial h}{\partial x}=0\right]$ or $\psi=\psi_o$

Figure 31. Two-dimensional mathematical model of Ross Creek Basin..

carefully so as to approximate a direction perpendicular to the regional trend, and parallel to the regional hydraulic gradient. It is presumed that the flow pattern in this cross section would be representative of the regional groundwater flow.

Essentially, the domain is made up of the hydrogeologic environment (Figure 30). The right vertical boundary which coincided with the crest of the Cypress Hills was assumed to be impermeable. Also, a bottom impermeable boundary was located at the contact of the Alberta Group and Milk River Formation. For the steady state runs, the top boundary was the water table, which was assumed to coincide with the surface topography because water level data were not enough to define it in the required detail. The justification for this assumption is that water level data indicated the water table to fall within 20 m from the surface in all cases. Furthermore, this assumption is documented elsewhere (Jones, 1960; Farvolden, 1961; Tóth, 1962). The left vertical boundary was assumed to be impermeable in some cases, and as a constant head boundary in others. The equation which governs groundwater flow through the domain is the steady state form of Richards' equation (Richards, 1931):

$$\frac{\partial}{\partial x}\left(K_x\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial z}\left(K_z\frac{\partial h}{\partial z}\right) = 0 \qquad (5)$$

where: $K_x$ = hydraulic conductivity of the medium in x direction [L/T];
$K_z$ = hydraulic conductivity of the medium in z direction [L/T];
$h$ = hydraulic head in the medium [L]; and
$x,z$ = space coordinates [L];

Equation (5), together with the domain and boundary conditions described above completed the boundary value problem that was solved.

The hydraulic head distribution was determined using two different finite element numerical models. First a computer model which solves (5) for the potential distribution in two-dimensional cross sections (Frind, 1971) was employed in the solution to the problem. The model uses the Galerkin procedure and Green's theorem to obtain elemental integral equations for the domain of interest, which are then solved by the Cholesky decomposition matrix algebra method. Details of the method of solution are presented in the papers by Frind (1971, 1982), Pinder and Frind (1972), and Frind and Matanga (1985). The top boundary was treated by specifying a hydraulic head value equal to the surface elevation at each node along the boundary. Several runs were processed varying the left boundary conditions and the hydraulic parameters (Kx and Ky) of the units comprising the aquifers and aquitard, in order to simulate measured hydraulic heads at the points indicated on Figure 32. Initially, field estimated hydraulic conductivities were used, and homogeneous and isotropic conditions were assumed. The aquifers were maintained at a hydraulic conductivity of two orders of magnitude higher than that of the aquitard throughout the simulations. Stepwise adjustments in hydraulic conductivities were made with reference to values obtained from literature. After each simulation, simulated hydraulic heads were compared with the measured point hydraulic heads. Changes were then made as necessary and simulations repeated. Measured point hydraulic heads were used for the calibration because no reliable hydraulic cross section could be constructed from the available water level data. A certain degree of doubt may be associated with results obtained in this manner, but it is felt that a range of flow patterns resulting from this analysis would be reliable. Best match between observed and simulated hydraulic heads was produced with the hydraulic conductivity values indicated in Table 8. The resulting hydraulic head distribution is displayed in Figure 32. The correlation between observed and simulated hydraulic heads is better at the lower topographic elevations than it is at higher elevations, probably due to the better coincidence of the surface topography with the water table at the lower elevations. Nevertheless, it is felt that the results of the mathematical model are in good correlation with the general nature of the observed hydraulic head distribution in the basin.

Having determined the hydraulic head distribution and regional hydraulic conductivities, it was necessary to estimate the flow pattern of groundwater. Also, the flow pattern would serve a second role of further calibration of the model. For this purpose, the mathematical model was reformulated in terms of stream functions.

A stream function is a function which has a constant value along a streamline, hence a streamline is a line which joins points of equal stream functions. Also, a streamline is a curve that is everywhere tangential to the specific discharge vector, thus it indicates the direction of flow at every point in the flow domain. Therefore, the calculation of stream functions leads to the construction of streamlines which indicate flow directions and facilitate groundwater discharge calculation as described in Bear (1979).

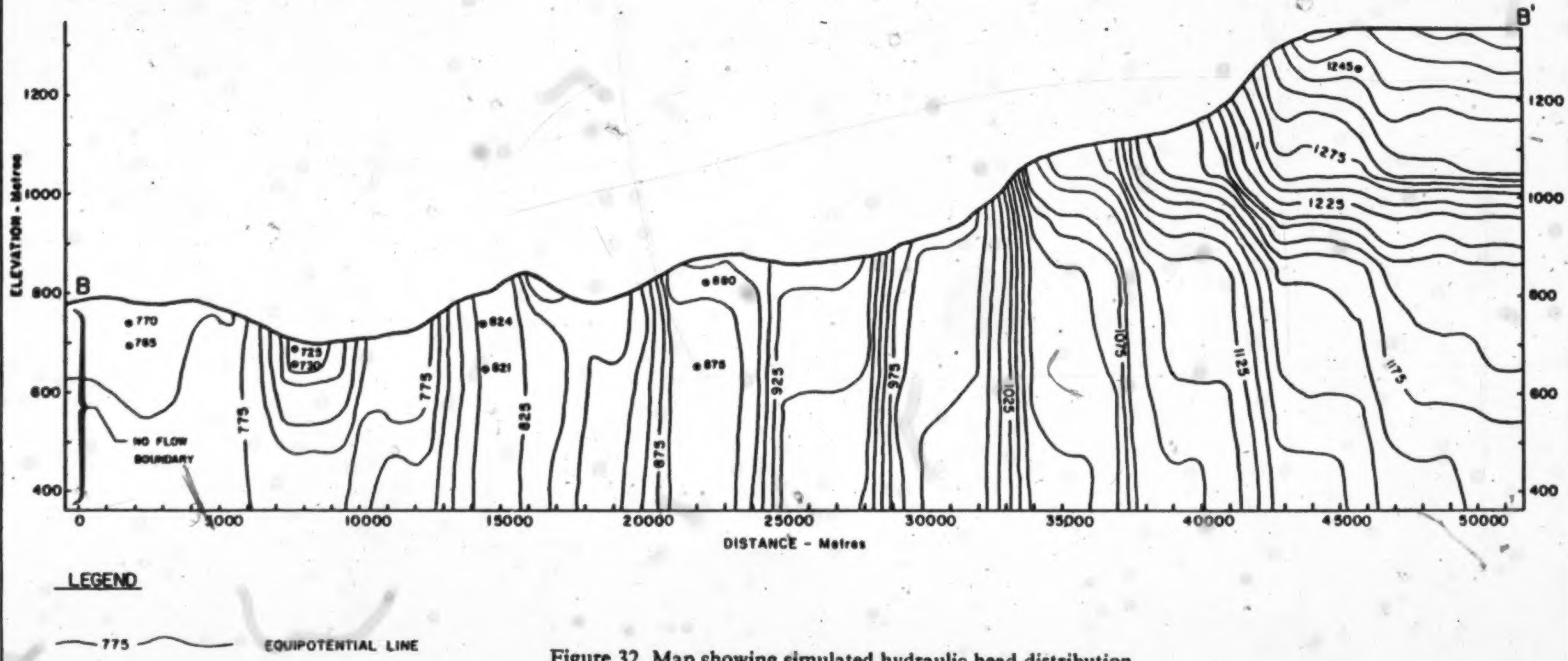


Figure 32. Map showing simulated hydraulic head distribution.

Table 8. Adjusted hydraulic conductivities used in numerical models.

| Formation | | Hydraulic conductivity (m/s) | |
|---|---|---|---|
| Aquifer | Aquitard | Horizontal (Kh) | Vertical (Kv) |
| | Cypress Hills | $8.4 \times 10^{-5}$ | $8.4 \times 10^{-7}$ |
| | Ravenscrag | $4.4 \times 10^{-5}$ | $4.4 \times 10^{-7}$ |
| | Frenchman | $4.5 \times 10^{-6}$ | $4.5 \times 10^{-8}$ |
| | Eastend | $4.5 \times 10^{-5}$ | $4.5 \times 10^{-7}$ |
| | Bearpaw | $4.2 \times 10^{-6}$ | $4.2 \times 10^{-8}$ |
| | Oldman/Foremost | $4.4 \times 10^{-5}$ | $4.4 \times 10^{-7}$ |
| | Pakowki | $4.2 \times 10^{-6}$ | $4.2 \times 10^{-8}$ |
| | Milk River | $8.4 \times 10^{-5}$ | $8.4 \times 10^{-7}$ |
| | Quaternary Drift | $2.1 \times 10^{-6}$ | $2.1 \times 10^{-8}$ |
| Quaternary Drift | | $4.2 \times 10^{-3}$ | $4.2 \times 10^{-5}$ |

The equation that describes the distribution of the stream functions in two dimensions is (Bear, 1979):

$$\frac{\partial}{\partial x}\left(\frac{1}{K_z}\frac{\partial\Psi}{\partial x}\right)+\frac{\partial}{\partial z}\left(\frac{1}{K_x}\frac{\partial\Psi}{\partial z}\right)=0 \qquad (6)$$

where: $\Psi$ is the stream function in the medium $[L^2/T]$.

Equation (6) was evaluated for Ross Creek Basin with the numerical model of Frind (1971). As the original model was written for the solution of potential dstribution only some modifications were necessary. The modifications include: 1. the replacement of Kx by 1/Kz and Kz by 1/Kx during the calculation of the elemental conductance matrix; and 2. the adjustment of the process by which the second-type boundary conditions which now replace the first-type boundary conditions (Frind and Matanga, 1985) of the potential formulation, are read into the program. Furthermore, to facilitate the large data input, and to expedite visual inspection of the results, the computer program was expanded to accomodate a grid generator and a plotting sub-routine. Thus, the results were displayed as a computer generated groundwater flow pattern map of the basin (Figure 33). The scale of Figure 33 has been exaggerated vertically but a direct interpretation of the flow pattern is possible.

Qualitatively, groundwater flow in Ross Creek Basin consists of local, intermediate and regional flow systems. Figure 33 shows that about 40 percent of the groundwater in the basin is recharged in the areas around the Cypress Hills. One-half of this water is discharged through a local flow system at a lower topographic elevation adjacent to the hills, and through an intermediate system into the Gros Ventre Creek valley, whereas the remaining one-half is discharged at the Ross Creek valley through a regional flow system. Figure 33 also indicates that the remaining 60 percent of the groundwater moving through the basin is recharged through local topographic highs on adjacent sides of Gros Ventre and Ross Creeks.

The calculated recharge-discharge picture of the flow pattern in the basin leads to a series of alternating recharge and discharge areas (Figure 33). A comparison of Figures 12 and 33 shows a good correlation between recharge and discharge areas of both figures, giving credit to the surficial mapping technique for groundwater reconnaissance studies in Ross Creek Basin. Furthermore, the coincidence of recharge and discharge areas is a further indication that the

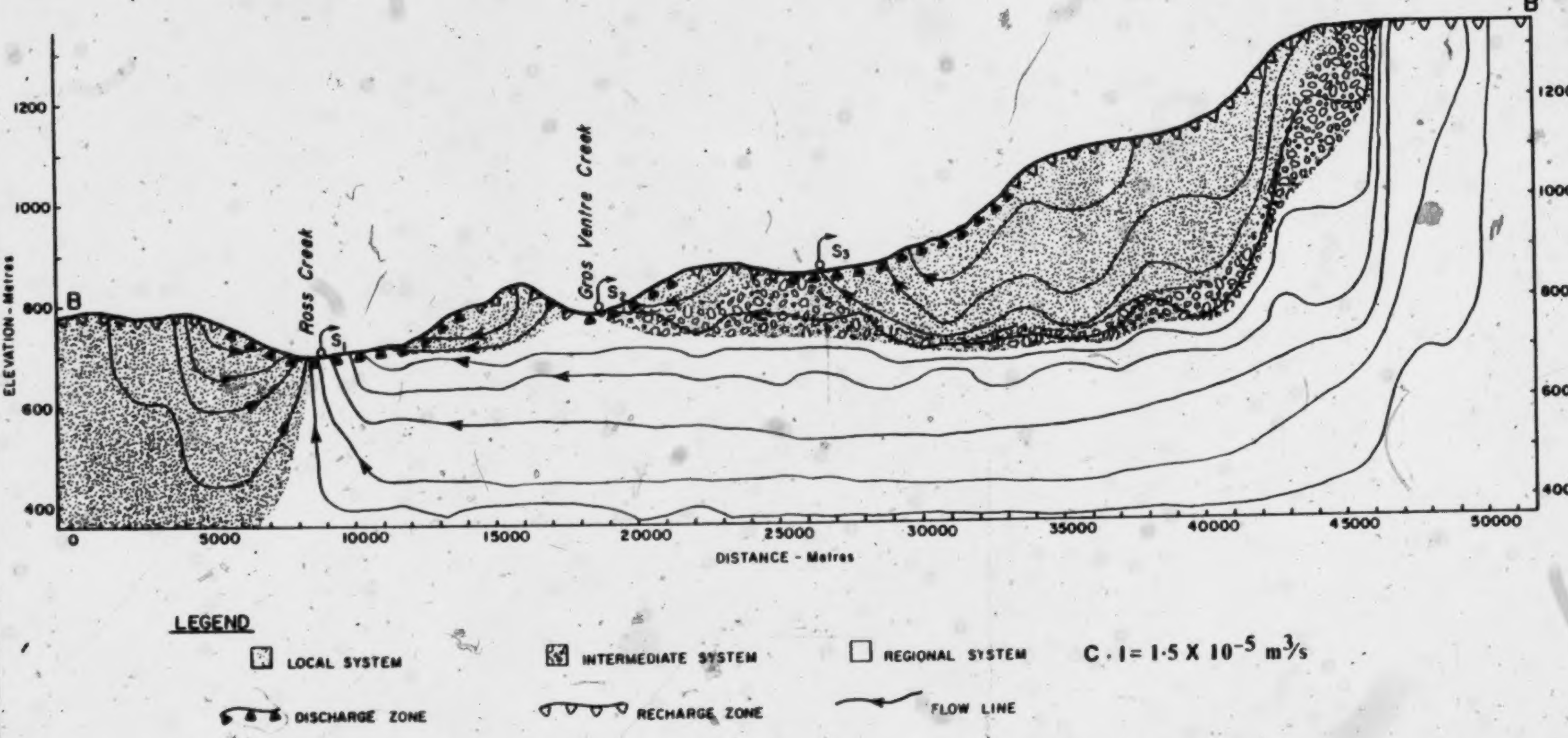

B
B'
Ross Creek
Gros Ventre Creek
$S_1$
$S_2$
$S_3$
ELEVATION - Metres
1200
1000
800
600
400
0
5000
10000
15000
20000
25000
30000
35000
40000
45000
50000
DISTANCE - Metres
LEGEND
LOCAL SYSTEM
INTERMEDIATE SYSTEM
REGIONAL SYSTEM
C · I = 1·5 X 10$^{-5}$ m$^3$/s
DISCHARGE ZONE
RECHARGE ZONE
FLOW LINE

Figure 33. Map showing simulated flow pattern.

flow pattern, and thus, the hydraulic head distribution determined by the models are representative of the basin's groundwater regime.

Quantitatively, Figure 33 can be analyzed easily to determine the total discharge through the system. Recognizing that the contour interval of Figure 33 is $1.5\times10^{-5}$ $m^3/s$, a total discharge (equal to recharge) of $2.0\times10^{-4}$ $m^3/s$ is estimated for the basin. This is the natural basin yield (Freeze, 1967; Freeze and Witherspoon, 1968) or the dynamic reserves (Tóth, 1982), the lower limit of groundwater that can be developed theoretically from Ross Creek Basin on a sustained basis. The calculated basin yield converts to 118 mm or 22 percent of the total precipitation of 531 mm. The similarity between the value of dynamic reserves obtained by this theoretical method and the value (113 mm) estimated from measured water level fluctuations is remarkable. However, if one recognizes that it is the same parameter being calculated by two different methods, their similarity could be expected.

The model by Frind (1971) was employed in the calibration process because it is cheap to execute, and a large number of runs was required to achieve a satisfactory calibration. However, this model is not equipped to handle the necessary transient simulations. To achieve this objective, all the calibrated parameters of the groundwater system were passed on to the model by Davis and Neuman (1983). This model is particularly adaptive to the present problem because it was necessary to trace the position of the water table with time. However, a plotting sub-routine was added to the program in order permit visual inspection of the temporal position of the water table during the simulations. Termed UNSAT2, the model has been developed over a number of years through a series of papers. The original documentation is given in Neuman et al. (1974) The theory is described fully in the papers by Neuman (1974, 1975). This model has been applied successfully in the solution of various flow problems (Feddes et al., 1974; Kroszynski and Dagan, 1975; Wei and Sheih, 1979; Zaslavsky and Sinai, 1981). The model treats the continuum of pressure heads that exists between the saturated and unsaturated zones simultaneously, thus, the position of the free surface can be defined without considering it to be a boundary within the system. For the region of flow, the model solves the pressure head equivalent of the general equation of transient flow through variably saturated porous media. Further details about the model may be obtained from Davis and Neuman (1983).

In the first instance, the model was executed once to redetermine the steady-state hydraulic head distribution. During this run, a pressure head value of zero was specified at each node along the surface topography to simulate the water table. The final steady-state hydraulic head distribution obtained by this model was similar to that of the previous model, with the largest discrepancy between nodal values being less than 1 m. Once it was obtained, the natural steady-state distribution of hydraulic heads served as the initial conditions for the transient analysis of the basin's groundwater regime that followed.

### 3.2.2 Response of the groundwater regime to development

The purpose of the transient analysis was dual: 1. to determine the best possible well location, its rate of production, and its role in meeting water demands in the basin; and 2. to determine the ultimate steady-state groundwater flow pattern and thus, advice on an appropriate management strategy.

To achieve this purpose, the model by Davis and Neuman (1983) was used to develop the basin's groundwater system. The initial conditions were those of the natural steady-state hydraulic head distribution (Figure 32). For the same reasons discussed in Ophori (1986, Figure 44), the unsaturated zone was characterized with data from Sophocleous (1978). Since the hydrogeologic environment (Figure 30) resembles the ATR and BTR environments (Ophori, 1986; Figure 22) of the Plains Regions in which the aquifer lies in the discharge area, the best position of a well in terms of maximum rates of production is easily identifiable. Only the aquifer at the lower Ross Creek valley is developed here because the other aquifer lies in the property of the Cypress Hills Provincial Park. Such a well should be located at the middle of the aquifer (Ophori, 1986; Figure 58). With the well located thus (point W, Figure 34), the model was run to steady state several times with various possible conditions of recharge from precipitation and available drawdown.

During the first run (run 1) a recharge rate of $3.0 \times 10^{-4}$ m/day, distributed evenly along the water table was allowed into the system to simulate the recharge rates estimated both theoretically (section 3.2.1) and through dynamic reserves calculations (section 2.8). A maximum permissible drawdown equal to two-third of the total available drawdown was assumed. To simulate pumping in the well, the water level in the well was held constant at a hydraulic head value equivalent to the permissible drawdown throughout the run. The model was then run until steady state was reached.

The interpretation of the results from the runs was based mainly on the findings of Ophori (1986) that: 1. the stability of a groundwater system under development is marked by positive values of transitional-basin-yield (TBY) and a steady water table; 2. the transition between stable and unstable conditions is associated with low negative, zero, or low positive TBY and a steady water table; and 3. instability is registered by high negative TBY and an unsteady water table. TBY is the cumulative inflow of water into a groundwater system due to development at a particular site, during the transient period between the initial natural steady-state conditions of flow and the new steady-state conditions resulting from the development.

Results of this run showed that although steady state was achieved eventually, the calculated TBY (Table 9) is negative indicating that the system is in the transition state, an unwanted situation. The results also showed that groundwater movement in the vicinities of springs $S_2$ and $S_3$ (observed during the hydrogeological mapping) remained upward at the end of the run. The water table still intersects the land surface in these areas, thus, groundwater is discharged continuously by $S_2$ and $S_3$. It is inferred from these results that, whereas enough water is recharged into the system to maintain groundwater discharge at $S_2$ and $S_3$, the hydraulic conductivity of the aquitard is lower than one needed to conduct enough water through the aquitard to the aquifer and thus to the well, in order to ensure the continuous stability of the system.

The situation is different in the area around $S_1$. The natural discharge area around $S_1$ is converted to a recharge area by pumping from the well. This is another detrimental effect of such a pumping scheme, as downstream water users will be deprived of their share of the natural supply of water along the creek.

The second run (run 2) was designed to represent a situation in which the recharge into the system is areally restricted; thereby simulating the entry of water into the system through recharge areas only, depression focussed recharge, and so on. During this run, the same recharge rate as in run 1 was concentrated at the nodes in the recharge areas along the water table, all other parameters being the same as in the first run. The results and interpretations thereof were similar in every respect to those in the preceding run, except that a little more negative TBY was calculated (Table 9). This is an indication that the system will be less stable

Table 9. TBY values for runs 1 to 5.

| Run | TBY ($m^3$) |
|---|---|
| 1 | -459 |
| 2 | -1 244 |
| 3 | -3 122 |
| 4 | -2 740* |
| 5 | 96 |

* no steady state attained

to development if the recharge were areally restricted.

In run 3, all the parameters of run 1 were used except that the recharge rate was reduced by one-half to $1.5 \times 10^{-4}$ m/day in order to simulate the conditions of a drier year. Once more, results similar to those of run 1 were obtained with an appreciably more negative TBY. In addition to the system being much less stable, an interesting deduction is that $S_2$ and $S_3$ continue to discharge groundwater. This situation suggests that if the actual groundwater recharge is half of the calculated value, most of the springs at the intermediate and high levels in Ross Creek Basin will continue to flow in the presence of development at the lower Ross Creek valley.

Run 4 simulated the conditions of a completely dry year, therefore, recharge into the system was zero, and all other parameters remained as they were in run 1. No steady state was obtained. Drawdown was slow but steady, and the run was stopped after enough time was allowed to obtain the required results. In this case, the water table fell below the ground level everywhere in the system (Figure 34). Thus, the development of the basin during a long period of no recharge into the system will cause unstable conditions, accompanied by the drying up of the springs.

Following these findings, several runs were processed to determine the optimal pumping plan for the basin. For each consecutive run, the permissible drawdown was reduced systematically and the value of TBY was examined until it became positive. In all these runs, the calculated recharge rate of $3.0 \times 10^{-4}$ m/day was used, and all other parameters corresponded to those of run 1. A positive value was first calculated in a run (run 5) in which the permissible drawdown was approximately equal to one-third of the available drawdown.

In run 5 which represents stable conditions of development in the basin, permissible drawdown is one-third of the available drawdown, springs $S_2$ and $S_3$ are discharging groundwater, and there is no discharge from $S_1$. This is viewed as the ultimate groundwater regime that will result from the optimal development of groundwater in Ross Creek Basin.

Another result obtained from the model calculations is the rate of pumping at the production well which corresponds to the optimal development. From run 5, the calculated pumping rate (obtained directly as model output) is $8.8 \times 10^{-5}$ m³/s. Field investigations indicate presently that the aquifer stretches for a length of approximately 1 000 m perpendicular to the modelled cross section. If this width is considered, together with a safety factor of 0.5, a total

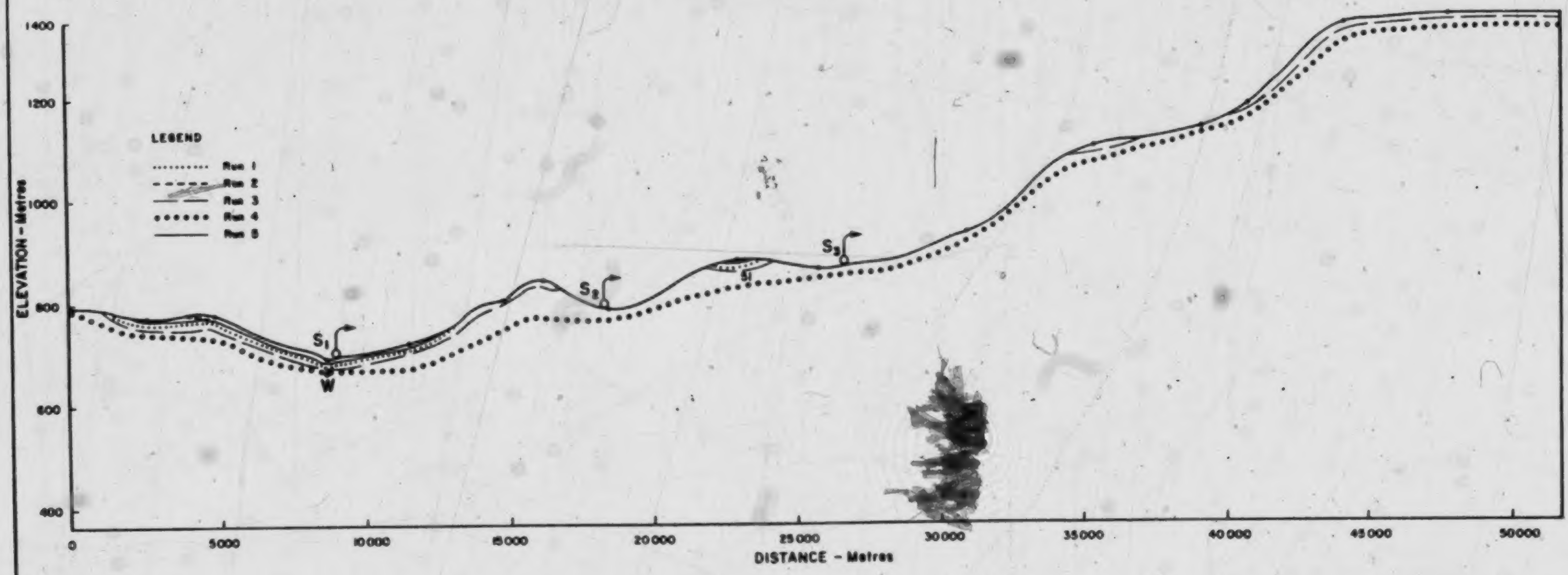


Figure 34. Position of water table for runs 1 to 5.

pumping rate of $4.4\text{x}10^{-2}$ $m^3$ /s is obtained.

Considering these theoretical results, the following deductions can be made about groundwater management in Ross Creek Basin: 1. the only aquifer of significance that was detected during the field studies in the basin has a potential production rate of $4.4\text{x}10^{-2}$ $m^3/s$; this rate is greater than the demand for total municipal supply ($6.8\text{x}10^{-3}$ $m^3/s$), and less than the demand for domestic ($5.8\text{x}10^{-2}$ $m^3/s$) and irrigation ($2.6\text{x}10^{-1}$ $m^3/s$) uses; 2. at the estimated safe rate of production, springs at intermediate and higher elevations will continue to discharge groundwater, thus, most of these springs can be licensed for use at their natural rates of discharge, without danger to the groundwater regime; and 3. at the calculated safe pumping rate, however, natural discharge into the lower Ross Creek is captured completely, thereby jeopardizing the water rights of downstream water users; the tradeoff between the advantages and disadvantages of this effect was not considered further in this study.

In order to satisfy the water rights of downstream users, that is, to allow some natural discharge into Ross Creek at the production site, a number of runs was processed with continuous reduction in the permissible drawdown. It was observed that pumping becomes negligible for this objective to be satisfied.

# 4 SUMMARY AND RECOMMENDATIONS

## 4.1 Summary

The topography of Ross Creek Basin varies from hills in the south, through a highly dissected knob-and-kettle topography "bench" area in the centre, to relatively flat low-lying land in the north. The effect of the stagnant puddles of water in the kettles on the groundwater regime is not known, but some physiographic features appear to wield a significant amount of influence on the groundwater in the area. For example, the water table resembles the surface topography, thus, the principal component of groundwater flow is from the southern highlands to the low areas in the north; and the soil type boundaries coincide fairly well with boundaries of recharge and discharge areas, suggesting that soil development in the area is associated with the groundwater flow pattern.

The area is underlain by a bedrock of alternating layers of Cretaceous marine shales and nonmarine sandstones. The shale units are mixed with siltstone and sandstone, whereas the sandstone layers are intercalated with siltstone and shale. This intermixing occurs in such proportions as to produce formations of approximately uniform sandstone, siltstone and shale contents. Occasionally, little pockets of clean sandstone occur. The glacial drift has been classified into morphostratigraphic units, purely on the basis of geographic position and inferred geologic history. The drift is thickest at the stream channels, ranging in thickness from 0 to 30 m. Till is the major component of the drift, but glacio-fluvial sand and sandy gravels, and glacio-lacustrine silt and clay occur frequently in places. In most cases, the glacio-fluvial sand and sandy gravels only produce groundwater for small household uses because of their restricted sizes. In only one case, a glacio-fluvial aquifer was considered for general water supply. In general, the bedrock units and till material possess hydraulic conductivities of $2.8 \times 10^{-10}$ to $2.8 \times 10^{-7}$ m/s, whereas the glacio-fluvial sand and sand gravels have higher hydraulic conductivities of up to $5.6 \times 10^{-4}$ m/s. Porosity probably range from 22 to 35 percent for the geologic materials of the basin.

Three periods of annual water cycle are experienced in the basin: a period of soil moisture utilization, accompanied by soil moisture deficit; a period of soil recharge; and a period of water surplus. Although annual precipitation exceeds annual actual evapotranspiration, no moisture surplus is available for groundwater recharge in a dry year.

Moisture deficits of 309 mm and 275 mm, mark dry and average years, respectively. In a wet year, however, moisture deficit may be as low as 216 mm, with a moisture surplus of 79 mm. It is evident therefore, that groundwater recharge occurs during average conditions and wet years only. These variations, in addition to possible monthly and daily variations are in support of a water management plan for the basin.

On the basis of the mapping of surficial manifestations of groundwater flow, the basin was sub-divided into area of downward, rectilinear, and upward flow, a helpful initial process in understanding the basin's groundwater regime.

Calculated hydraulic conductivities and transmissivities range between $4.6 \times 10^{-7}$ to $8.4 \times 10^{-6}$ m/s, and $5.6 \times 10^{-6}$ to $1.4 \times 10^{-3}$ $m^2/s$, respectively. Only a few local bodies of rock possess high transmissivities (greater than $1.0 \times 10^{-4}$ $m^2/s$). In combining these transmissivities with available drawdown to estimate twenty-year safe yields, only two relatively small aquifers ($Q_{20}$ greater than $7.6 \times 10^{-4}$ $m^3/s$) are identified, one of which is a property of the Cypress Hills Provincial Park. The remaining portion of the basin is an aquitard.

The range of mineralogical compositions of the rocks in recharge areas, and operating geochemical processes seem to smooth out geochemical constituents along groundwater flow paths. Areas of high total dissolved solids coincide with those of high sulphate and low bicarbonate and carbonate, with the converse being also true, in most cases. The occurrence of an appreciable amount of chloride around point C (Figure 21) seems to identify this area as a significant discharge area, containing "old" water, and conforms with the last stages of the Cheboratev sequence (Cheboratev, 1955). On the other hand, the existence of a wide range of calcium to magnesium ratio may be associated with variation in original calcium to magnesium ratios in the host rocks, although their trends of occurrence are in agreement with inferred groundwater flow directions. Calcium-magnesium-sulphate-bicarbonate and sodium-sulphate-bicarbonate facies are the main "discharge facies", and Ca-Mg-$HCO_3$ and Na-$HCO_3$ facies are in close association with recharge and midline areas. Generally speaking, hydrochemical facies are in good correlation with the groundwater flow pattern, and more than 50 percent of the basin's groundwater is potable.

Analysis of groundwater level fluctuations indicates that 113 mm or 21 percent of the annual precipitation is recharged and discharged through the basin. This is presumed to be the source of water to the ever discharging springs of the basin. A sustainable groundwater production rate of 1.1 m³/s is inferred by these results. However, the effect of variation in the geologic framework on the distribution of the rate of production across the basin is not considered in the calculation of the dynamic reserves.

The disparity between groundwater quantities estimated by water balance calclations, and the methods of dynamic reserves and numerical simulation is worth mentioning here. A detailed water balance study of the Cypress Hills area was undertaken by Wong (1974) who estimated a water surplus of zero for most of the northern parts of the basin, indicating that groundwater recharge may not occur at all. Most casual observers would agree that the prospects for significant amount of recharge in the basin is poor because of the sparse cover of drought-resistant grasses and shrubs, and mean annual actual evapotranspiration rates that approach or exceed mean annual precipitation (Laycock, 1967). On the other hand, the methods of dynamic reserves and numerical simulation estimate groundwater recharge of 113 mm and 118 mm, respectively. Whereas it possible that the latter methods may have overestimated the groundwater recharge because the basin does not appear to be sufficiently wet, the former method appears to underestimate groundwater recharge for the following reasons: 1. the continuous discharge from springs in the basin must be maintained by a source of water; 2. fluctuations in water levels in wells of the area indicate a source of supply; 3. the water table is sub-parallel to the surface topography, a condition that may not exist in the absence of groundwater recharge; 4. phreatophyte patches are found in several zones in the basin, and a perennial tributary to Ross Creek has been formed from discharge by springs at one site (SW-14-12-4W4); 5. a water surplus of up to 152 mm has been estimated at the Cypress Hills Plateaux (Wong, 1974), the area that is believed to be the main zone of recharge of water into the basin in the present study; and 6. groundwater recharge values of up to 20 percent have been observed in New Mexico (Stephens, 1985), in areas where mean annual evapotranspiration is several times greater than mean annual precipitation. A determination of the actual quantitative relationship between rates of groundwater discharge and the features above is beyond the scope of this report. Until such a relationship is quantifiable, the disparity in the estimates of groundwater recharge by these methods cannot be reconciled.

Surface runoff from the basin is estimated as 3.8 percent of the annual precipitation. Most of the runoff (70 percent) occurs in the spring months of March and April, and it is zero during most months of the year. In general, runoff is unpredictable and requires that water be stored during floods for use in the time of need.

As evidenced by applications for water licenses, water is needed for three major purposes in Ross Creek Basin. These are irrigation ($2.6 \times 10^{-1}$ m$^3$/s), domestic use ($5.8 \times 10^{-2}$ m$^3$/s), and municipal purpose ($6.8 \times 10^{-3}$ m$^3$/s).

The estimated total streamflow (0.4 m$^3$/s) and groundwater flow through the basin (1.1 m$^3$/s) is greater than the total demand for water ($3.3 \times 10^{-1}$ m$^3$/s). The problem is that of the temporal and spatial distribution of the water resources as evidenced by streamflow and hydrometeorological calculations. Under these conditions, the role of the subsurface is to provide space for water storage during floods for release and use in the time of need; that is, aquifers are required that can be recharged either artificially or through induced infiltration. Such aquifers are not known to occur in the basin presently from a knowledge of the geology and aquifer hydraulics. Furthermore, an analysis of dynamic reserves shows that much water is being recharged into the groundwater system already, but the lack of the appropriate site for its exploitation has left this water to discharge and evaporate through numerous springs, and directly into the atmosphere. Strictly speaking, the groundwater system is not connected directly with the surface water, therefore it has to be operated separately. Groundwater management in the basin is a question of how best it can be exploited and for what use. This is a problem of an understanding of the system's response to development and has been treated as such.

Twenty-year safe yield analysis indicates that a maximum of $4.1 \times 10^{-3}$ m$^3$/s can be produced from a restricted locality in the basin to mine the groundwater completely in 20 years. These tests are short and do not reflect the long-term response of the system to development. The production rate estimated by this method is based mainly on a calculation of aquifer parameters, mainly transmissivity. The existence of high transmissivities alone does not guarantee high safe rates of production (Ophori, 1986); there must be a source of water into the system to be produced at a later date. In fact, the higher the transmissivity, the faster the groundwater is mined from the system, in the presence of insufficient recharge. The recharge-discharge regimes of the groundwater system must be considered in a basin-wide study that is designed to produce a licensing plan for its springs as is the case herein. Aquifer hydraulics does not provide for such considerations.

The formulation of the model for the transient analysis of the groundwater system in Ross Creek Basin accounts for some shortfalls of aquifer hydraulics. This model shows that

$4.4\times10^{-2}$ $m^3/s$ of groundwater can be pumped from the basin on sustained basis. This rate of production does not affect the natural discharge of groundwater from the springs at intermediate and higher topographic elevations in the basin. Therefore, these springs may be developed for use even while groundwater is being pumped. The springs may be licensed for use in instances where they do not form the main source of water supply to downstream water users, without danger to the groundwater system.

The groundwater produced at this rate is enough to meet municipal water demands, or half of the domestic demands in the basin. However, the cost of production and transfer to the sites of need have not been considered in the study. Another unwanted effect of the groundwater pumpage is the obliteration of the springs ($S_1$) at the lower Ross Creek valley (the site of production). These springs are the sole source of water supply to water users downstream of the pumping site. The advantages or disadvantages of this effect have also not been analyzed in this study. An attempt to pump the system at a rate that would ensure the continued natural discharge of water to downstream users resulted in a pumping rate of approximately zero, suggesting that the system be left to operate under natural conditions.

In this study, the supply of water to users downstream from the proposed pumping site is important, thus, it is concluded that groundwater pumpage is an insignificant source of water supply to a centrally operated water system in the basin.

### 4.2 Recommendations

Regarding groundwater resources in Ross Creek Basin, the following recommendations are made:

1. All the springs in the basin should be surveyed to determine those that do not contribute to the water supply of users downstream from their points of discharge. Licenses may be issued for the development and use of such springs by their local owners.

2. A study should be conducted to determine the actual amount of water required by users along Ross Creek downstream from SW-14-12-4W4.

3. The Ross Creek valley should be dredged for a length of about 1 000 m with its center at SW-14-12-4-W4, to enhance the natural discharge by springs into the creek in this area. After satisfying downstream demands, the remaining water should be used to meet part of the basin's water demands.

4. Should the groundwater pumping option at SW-14-12-4-W4 be considered, an economic analysis of the cost and benefit of this action in terms of satisfying downstream demands, and meeting the basin's water demands partially should be conducted. A local and more detailed study of the area should be conducted to determine the best location of the pumping well.

5. Should any other groundwater study be planned for the basin, it should be aimed at an exploration for surficial aquifers that are possibly connected to Ross Creek hydraulically. In the event of a discovery of a suitable aquifer, an induced infiltration program should be planned.

6. In the absence of an adequate aquifer for groundwater storage as it is presently, surface storage of stream runoff during the spring months for use in the dry months appears to be a good option.

7. The present individual, small-scale pumpage of groudwater by farmers from local surficial sand and sandy gravels is in accord with the basin's groundwater regime and should be encouraged. At present, individual pumping of groundwater for household use appears to be the only role of the basin's groundwater.

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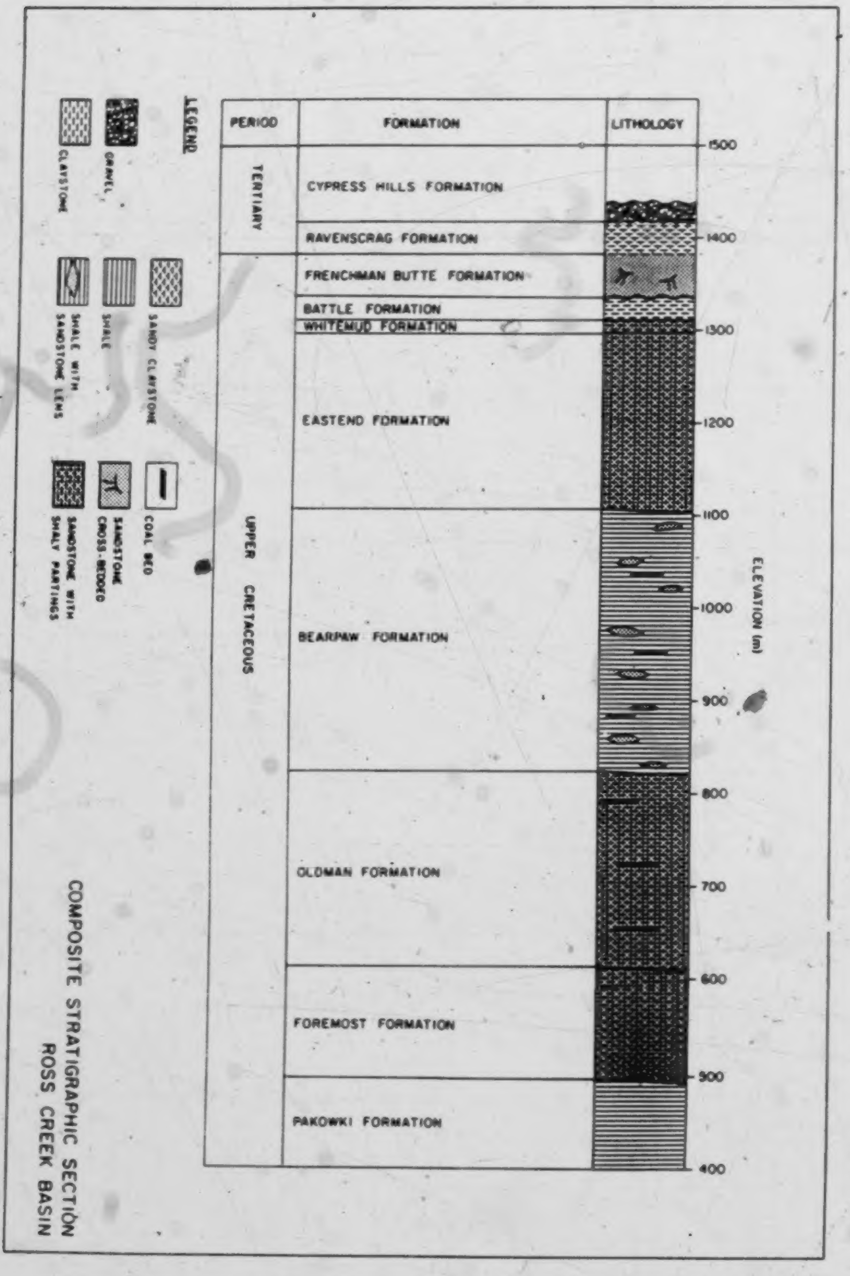
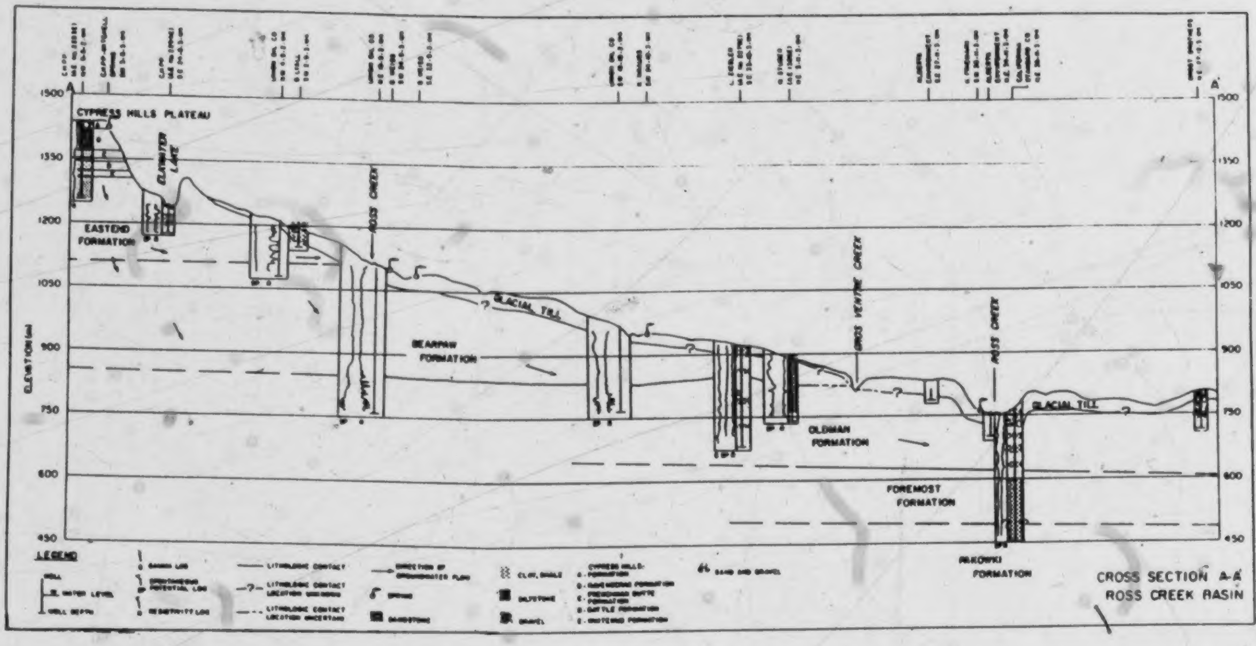
## Appendix 1 - Composite stratigraphic section and generalized geologic cross sections of Ross Creek Basin

(Traces on Figure 3)

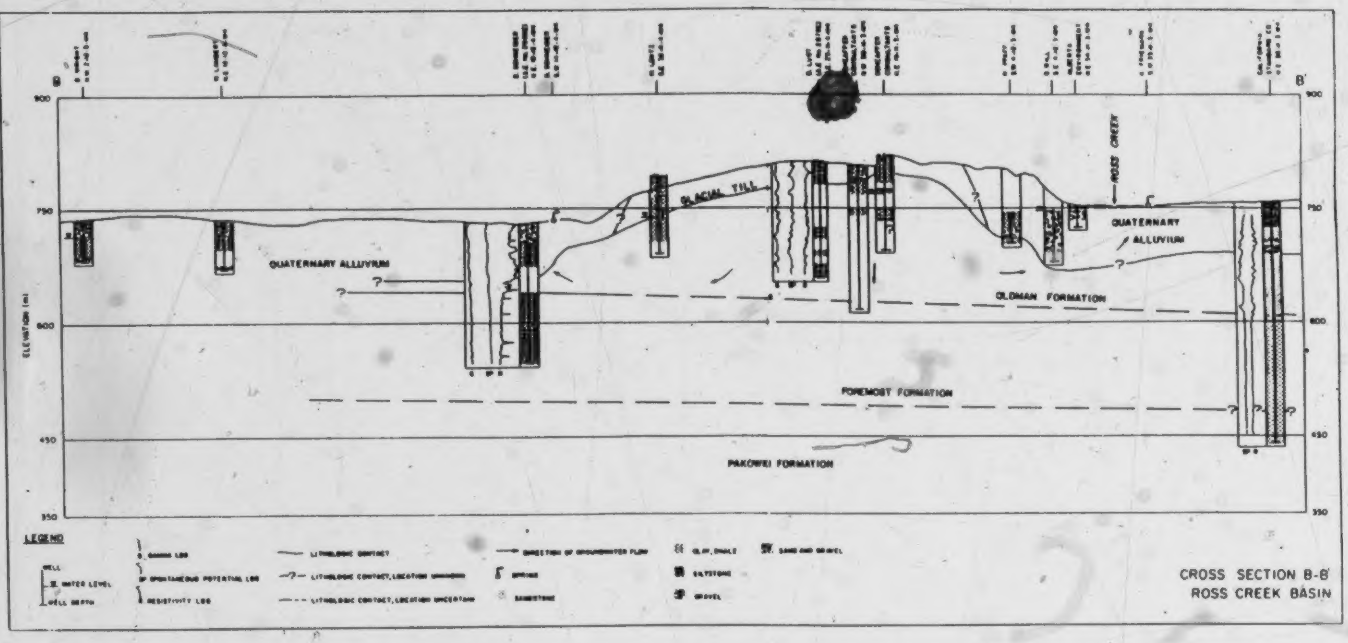
(source - Freeman, 1986)


LEGEND
GRAVEL
CLAYSTONE
SANDY CLAYSTONE
SHALE
SHALE WITH SANDSTONE LENS
COAL BED
SANDSTONE CROSS-BEDDED
SANDSTONE WITH SHALY PARTINGS
PERIOD
FORMATION
LITHOLOGY
TERTIARY
UPPER CRETACEOUS
CYPRESS HILLS FORMATION
RAVENSCRAG FORMATION
FRENCHMAN BUTTE FORMATION
BATTLE FORMATION
WHITEMUD FORMATION
EASTEND FORMATION
BEARPAW FORMATION
OLDMAN FORMATION
FOREMOST FORMATION
PAKOWKI FORMATION
ELEVATION (m)
1500
1400
1300
1200
1100
1000
900
800
700
600
500
400

COMPOSITE STRATIGRAPHIC SECTION
ROSS CREEK BASIN

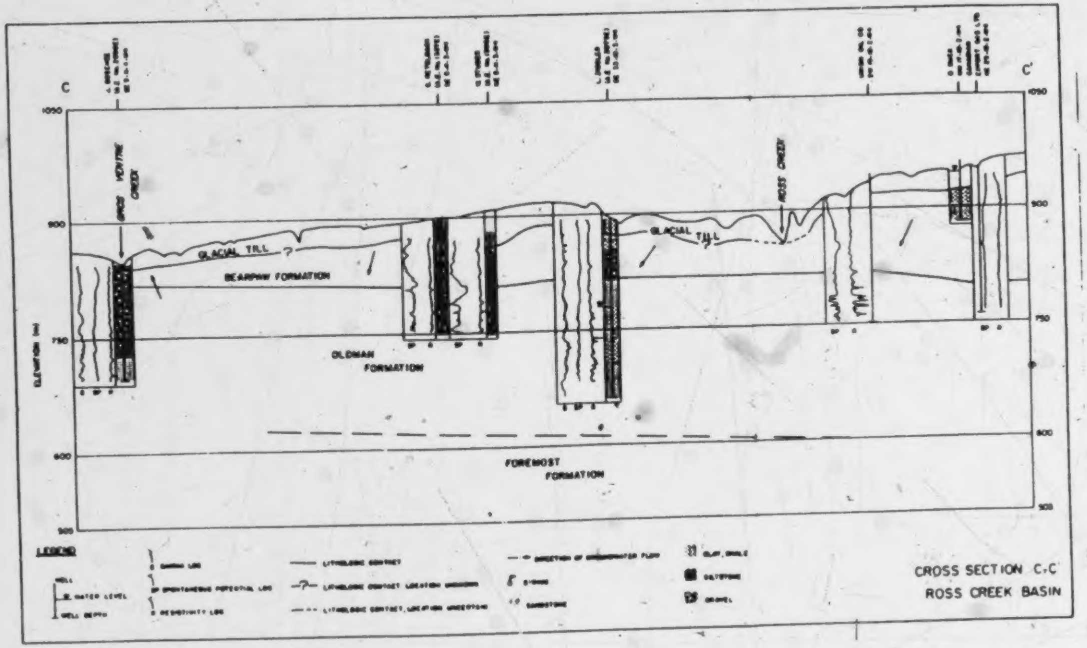

A
A'
1500
1350
1200
1050
900
750
600
450
ELEVATION (m)
CYPRESS HILLS PLATEAU
ELKWATER LAKE
EASTEND FORMATION
ROSS CREEK
GLACIAL TILL
BEARPAW FORMATION
GROS VENTRE CREEK
OLDMAN FORMATION
ROSS CREEK
GLACIAL TILL
FOREMOST FORMATION
PAKOWKI FORMATION
LEGEND
WELL
WATER LEVEL
WELL DEPTH
GAMMA LOG
SPONTANEOUS POTENTIAL LOG
RESISTIVITY LOG
LITHOLOGIC CONTACT
LITHOLOGIC CONTACT LOCATION UNKNOWN
LITHOLOGIC CONTACT LOCATION UNCERTAIN
DIRECTION OF GROUNDWATER FLOW
SPRING
SANDSTONE
CLAY, SHALE
SILTSTONE
GRAVEL
A - CYPRESS HILLS FORMATION
B - RAVENSCRAG FORMATION
C - FRENCHMAN BUTTE FORMATION
D - BATTLE FORMATION
E - WHITEMUD FORMATION
SAND AND GRAVEL

CROSS SECTION A-A'
ROSS CREEK BASIN



B
B'
900
750
600
450
350
ELEVATION (m)
ROSS CREEK
GLACIAL TILL
QUATERNARY ALLUVIUM
QUATERNARY
ALLUVIUM
OLDMAN FORMATION
FOREMOST FORMATION
PAKOWKI FORMATION
?
LEGEND
WELL
WATER LEVEL
WELL DEPTH
G GAMMA LOG
SP SPONTANEOUS POTENTIAL LOG
R RESISTIVITY LOG
LITHOLOGIC CONTACT
LITHOLOGIC CONTACT, LOCATION UNKNOWN
LITHOLOGIC CONTACT, LOCATION UNCERTAIN
DIRECTION OF GROUNDWATER FLOW
SPRING
SANDSTONE
CLAY, SHALE
SILTSTONE
GRAVEL
SAND AND GRAVEL

CROSS SECTION B-B'
ROSS CREEK BASIN


C
C'
1050
900
750
600
500
ELEVATION (m)
GROS VENTRE CREEK
ROSS CREEK
GLACIAL TILL
BEARPAW FORMATION
OLDMAN FORMATION
FOREMOST FORMATION
LEGEND
WELL
WATER LEVEL
WELL DEPTH
LITHOLOGIC CONTACT
SP
R

CROSS SECTION C-C'
ROSS CREEK BASIN

# Appendix 2 - Soil water balance calculations

YEAR: 1971

| | J | F | M | A | M | J | J | A | S | O | N | D | ANNUAL TOTAL |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| MEAN AIR TEMP. (T°C) | -14.7 | - 5.7 | - 2.8 | 6.8 | 13.2 | 16.1 | 18.4 | 22.0 | 11.6 | 5.9 | - 1.2 | -14.0 | |
| HEAT INDEX (I) | 0 | 0 | 0 | 1.59 | 4.35 | 5.87 | 7.19 | 10.08 | 3.58 | 1.29 | 0 | 0 | 33.95 |
| UNADJUSTED P.E. | 0 | 0 | 0 | 33 | 66 | 81 | 93 | 116 | 56 | 29 | 0 | 0 | |
| P.E. (mm) | 0 | 0 | 0 | 38 | 88 | 110 | 127 | 145 | 61 | 26 | 0 | 0 | 595 |
| P. (mm) | 43 | 7 | 16 | 9 | 46 | 41 | 9 | 21 | 27 | 18 | 9 | 13 | 259 |
| ST (12mm) | 12 | 12 | 12 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 9 | 12 | |
| AST. (mm) | 0 | 0 | 0 | -12 | 0 | 0 | 0 | 0 | 0 | 0 | 9 | 3 | |
| A.E. (mm) | 0 | 0 | 0 | 21 | 46 | 41 | 9 | 21 | 27 | 18 | 0 | 0 | |
| SURPLUS (mm) | 43 | 7 | 16 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 10 | 83 |
| DEFICIT (mm) | 0 | 0 | 0 | 17 | 42 | 69 | 118 | 124 | 34 | 8 | 0 | 0 | 412 |
| ST (50mm) | 50 | 50 | 50 | 21 | 0 | 0 | 0 | 0 | 0 | 0 | 9 | 22 | |
| AST. (mm) | 21 | 0 | 0 | -29 | -21 | 0 | 0 | 0 | 0 | 0 | 9 | 13 | |
| A.E. (mm) | 0 | 0 | 0 | 38 | 67 | 41 | 9 | 21 | 27 | 18 | 0 | 0 | |
| SURPLUS (mm) | 22 | 7 | 16 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 45 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 21 | 69 | 118 | 124 | 34 | 8 | 0 | 0 | 374 |
| ST (100mm) | 72 | 79 | 95 | 66 | 24 | 0 | 0 | 0 | 0 | 0 | 9 | 22 | |
| AST. (mm) | 43 | 7 | 16 | -29 | -42 | -24 | 0 | 0 | 0 | 0 | 9 | 13 | |
| A.E. (mm) | 0 | 0 | 0 | 38 | 88 | 65 | 9 | 21 | 27 | 18 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 45 | 118 | 124 | 34 | 8 | 0 | 0 | 329 |
| ST (150mm) | 72 | 79 | 95 | 66 | 24 | 0 | 0 | 0 | 0 | 0 | 9 | 22 | |
| AST. (mm) | 43 | 7 | 16 | -29 | -42 | -24 | 0 | 0 | 0 | 0 | 9 | 13 | |
| A.E. (mm) | 0 | 0 | 0 | 38 | 88 | 65 | 9 | 21 | 27 | 18 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 45 | 118 | 124 | 34 | 8 | 0 | 0 | 329 |
| ST (250mm) | 72 | 79 | 95 | 66 | 24 | 0 | 0 | 0 | 0 | 0 | 9 | 22 | |
| AST. (mm) | 43 | 7 | 16 | -29 | -42 | -24 | 0 | 0 | 0 | 0 | 9 | 13 | |
| A.E. (mm) | 0 | 0 | 0 | 38 | 88 | 65 | 9 | 21 | 27 | 18 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 45 | 118 | 124 | 34 | 8 | 0 | 0 | 329 |

YEAR: 1972

| | J | F | M | A | M | J | J | A | S | O | N | D | ANNUAL TOTAL |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| MEAN AIR TEMP. (T°C) | -16.1 | -12.9 | - 0.9 | 5.7 | 12.9 | 18.1 | 17.3 | 21.0 | 9.5 | 4.0 | - 9.1 | -17.5 | |
| HEAT INDEX (I) | 0 | 0 | 0 | 1.22 | 4.20 | 7.01 | 6.55 | 8.78 | 2.64 | 0.71 | 0 | 0 | 31.11 |
| UNADJUSTED P.E. | 0 | 0 | 0 | 30 | 66 | 93 | 89 | 107 | 42 | 21 | 0 | 0 | |
| P.E. (mm) | 0 | 0 | 0 | 35 | 88 | 126 | 122 | 134 | 52 | 19 | 0 | 0 | 576 |
| P. (mm) | 22 | 23 | 8 | 20 | 40 | 41 | 44 | 33 | 48 | 20 | 4 | 30 | 333 |
| ST (12mm) | 12 | 12 | 12 | 0 | 0 | 0 | 0 | 0 | 0 | 1 | 5 | 12 | |
| AST. (mm) | 0 | 0 | 0 | -12 | 0 | 0 | 0 | 0 | 0 | 1 | 4 | 7 | |
| A.E. (mm) | 0 | 0 | 0 | 32 | 40 | 41 | 44 | 33 | 48 | 19 | 0 | 0 | |
| SURPLUS (mm) | 22 | 23 | 8 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 23 | 63 |
| DEFICIT (mm) | 0 | 0 | 0 | 3 | 48 | 85 | 78 | 101 | 4 | 0 | 0 | 0 | 319 |
| ST (50mm) | 44 | 50 | 50 | 35 | 0 | 0 | 0 | 0 | 0 | 1 | 5 | 35 | |
| AST. (mm) | 22 | 6 | 0 | -15 | -35 | 0 | 0 | 0 | 0 | 1 | 4 | 30 | |
| A.E. (mm) | 0 | 0 | 0 | 35 | 75 | 41 | 44 | 33 | 48 | 19 | 0 | 0 | |
| SURPLUS (mm) | 0 | 17 | 8 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 25 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 13 | 85 | 78 | 101 | 4 | 0 | 0 | 0 | 281 |
| ST (100mm) | 44 | 67 | 75 | 60 | 12 | 0 | 0 | 0 | 0 | 1 | 5 | 35 | |
| AST. (mm) | 22 | 23 | 8 | -15 | -48 | -12 | 0 | 0 | 0 | 1 | 4 | 30 | |
| A.E. (mm) | 0 | 0 | 0 | 35 | 88 | 53 | 44 | 33 | 48 | 19 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 73 | 78 | 101 | 4 | 0 | 0 | 0 | 256 |
| ST (150mm) | 44 | 67 | 75 | 60 | 12 | 0 | 0 | 0 | 0 | 1 | 5 | 35 | |
| AST. (mm) | 22 | 23 | 8 | -15 | -48 | -12 | 0 | 0 | 0 | 1 | 4 | 30 | |
| A.E. (mm) | 0 | 0 | 0 | 35 | 88 | 53 | 44 | 33 | 48 | 19 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 73 | 78 | 101 | 4 | 0 | 0 | 0 | 256 |
| ST (250mm) | 44 | 67 | 75 | 60 | 12 | 0 | 0 | 0 | 0 | 1 | 5 | 35 | |
| AST. (mm) | 22 | 23 | 8 | -15 | -48 | -12 | 0 | 0 | 0 | 1 | 4 | 30 | |
| A.E. (mm) | 0 | 0 | 0 | 35 | 88 | 53 | 44 | 33 | 48 | 19 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 73 | 78 | 101 | 4 | 0 | 0 | 0 | 256 |

YEAR: 1973

| | J | F | M | A | M | J | J | A | S | O | N | D | ANNUAL TOTAL |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| MEAN AIR TEMP. (T°C) | - 6.3 | - 6.9 | 3.3 | 4.1 | 13.1 | 16.7 | 19.8 | 19.7 | 12.9 | 8.1 | - 8.7 | - 6.7 | |
| HEAT INDEX (I) | 0 | 0 | 0.53 | 0.74 | 4.30 | 6.21 | 8.03 | 7.97 | 4.20 | 2.08 | 0 | 0 | 34.06 |
| UNADJUSTED P.E. | 0 | 0 | 15 | 19 | 65 | 81 | 100 | 100 | 63 | 30 | 0 | 0 | |
| P.E. (mm) | 0 | 0 | 15 | 22 | 86 | 110 | 137 | 125 | 67 | 36 | 0 | 0 | 598 |
| P. (mm) | 2 | 9 | 1 | 31 | 6 | 124 | 7 | 37 | 20 | 17 | 24 | 19 | 297 |
| ST (12mm) | 12 | 12 | 0 | 9 | 0 | 12 | 0 | 0 | 0 | 0 | 12 | 12 | |
| AST. (mm) | 0 | 0 | -12 | 9 | -9 | 12 | -12 | 0 | 0 | 0 | 12 | 0 | |
| A.E. (mm) | 0 | 0 | 13 | 22 | 15 | 110 | 19 | 37 | 20 | 17 | 0 | 0 | |
| SURPLUS (mm) | 2 | 9 | 0 | 0 | 0 | 2 | 0 | 0 | 0 | 0 | 12 | 19 | 36 |
| DEFICIT (mm) | 0 | 0 | 2 | 0 | 71 | 0 | 118 | 88 | 47 | 19 | 0 | 0 | 345 |
| ST (50mm) | 37 | 46 | 32 | 41 | 0 | 14 | 0 | 0 | 0 | 0 | 24 | 43 | |
| AST. (mm) | 2 | 9 | -14 | 9 | -41 | 14 | -14 | 0 | 0 | 0 | 24 | 19 | |
| A.E. (mm) | 0 | 0 | 15 | 22 | 47 | 110 | 21 | 37 | 20 | 17 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 39 | 0 | 116 | 88 | 47 | 19 | 0 | 0 | 309 |
| ST (100mm) | 37 | 46 | 32 | 41 | 0 | 14 | 0 | 0 | 0 | 0 | 24 | 43 | |
| AST. (mm) | 2 | 9 | -14 | 9 | -41 | 14 | -14 | 0 | 0 | 0 | 24 | 19 | |
| A.E. (mm) | 0 | 0 | 15 | 22 | 47 | 110 | 21 | 37 | 20 | 17 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 39 | 0 | 116 | 88 | 47 | 19 | 0 | 0 | 309 |
| ST (150mm) | 37 | 46 | 32 | 41 | 0 | 14 | 0 | 0 | 0 | 0 | 24 | 43 | |
| AST. (mm) | 2 | 9 | -14 | 9 | -41 | 14 | -14 | 0 | 0 | 0 | 24 | 19 | |
| A.E. (mm) | 0 | 0 | 15 | 22 | 47 | 110 | 21 | 37 | 20 | 17 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 39 | 0 | 116 | 88 | 47 | 19 | 0 | 0 | 309 |
| ST (250mm) | 37 | 46 | 32 | 41 | 0 | 14 | 0 | 0 | 0 | 0 | 24 | 43 | |
| AST. (mm) | 2 | 9 | -14 | 9 | -41 | 14 | -14 | 0 | 0 | 0 | 24 | 19 | |
| A.E. (mm) | 0 | 0 | 15 | 22 | 47 | 110 | 21 | 37 | 20 | 17 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 39 | 0 | 116 | 88 | 47 | 19 | 0 | 0 | 309 |

YEAR: 1974

| | J | F | M | A | M | J | J | A | S | O | N | D | ANNUAL TOTAL |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| MEAN AIR TEMP. (T°C) | -12.4 | - 4.0 | - 3.8 | 8.0 | 9.6 | 18.7 | 20.7 | 16.3 | 11.8 | 9.2 | 0.5 | - 2.3 | |
| HEAT INDEX (I) | 0 | 0 | 0 | 2.04 | 2.69 | 7.37 | 8.59 | 5.98 | 3.67 | 2.52 | 0.03 | 0 | 32.89 |
| UNADJUSTED P.E. | 0 | 0 | 0 | 40 | 40 | 96 | 106 | 83 | 60 | 46 | 0 | 0 | |
| P.E. (mm) | 0 | 0 | 0 | 46 | 64 | 131 | 148 | 104 | 64 | 42 | 0 | 0 | 599 |
| P. (mm) | 14 | 18 | 28 | 45 | 53 | 45 | 23 | 48 | 3 | 5 | 1 | 8 | 291 |
| ST (12mm) | 12 | 12 | 12 | 11 | 0 | 0 | 0 | 0 | 0 | 0 | 1 | 9 | |
| AST. (mm) | 0 | 0 | 0 | -1 | -11 | 0 | 0 | 0 | 0 | 0 | 1 | 8 | |
| A.E. (mm) | 0 | 0 | 0 | 46 | 64 | 45 | 23 | 48 | 3 | 5 | 0 | 0 | |
| SURPLUS (mm) | 14 | 18 | 28 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 91 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 86 | 127 | 56 | 61 | 37 | 0 | 0 | 362 |
| ST (50mm) | 50 | 50 | 50 | 49 | 38 | 0 | 0 | 0 | 0 | 0 | 1 | 9 | |
| AST. (mm) | 7 | 0 | 0 | -1 | -11 | -38 | 0 | 0 | 0 | 0 | 1 | 8 | |
| A.E. (mm) | 0 | 0 | 0 | 46 | 64 | 83 | 23 | 48 | 3 | 5 | 0 | 0 | |
| SURPLUS (mm) | 7 | 18 | 28 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 53 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 48 | 122 | 56 | 61 | 37 | 0 | 0 | 324 |
| ST (100mm) | 57 | 75 | 100 | 99 | 88 | 2 | 0 | 0 | 0 | 0 | 1 | 9 | |
| AST. (mm) | 14 | 18 | 25 | -1 | -11 | -86 | -2 | 0 | 0 | 0 | 1 | 8 | |
| A.E. (mm) | 0 | 0 | 0 | 46 | 64 | 131 | 25 | 48 | 3 | 37 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 3 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 3 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 120 | 56 | 61 | 37 | 0 | 0 | 274 |
| ST (150mm) | 57 | 75 | 103 | 102 | 91 | 5 | 0 | 0 | 0 | 0 | 1 | 9 | |
| AST. (mm) | 14 | 18 | 28 | -1 | -11 | -86 | -5 | 0 | 0 | 0 | 1 | 8 | |
| A.E. (mm) | 0 | 0 | 0 | 46 | 64 | 131 | 28 | 48 | 3 | 5 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 117 | 56 | 61 | 37 | 0 | 0 | 271 |
| ST (250mm) | 57 | 75 | 103 | 102 | 91 | 5 | 0 | 0 | 0 | 0 | 1 | 9 | |
| AST. (mm) | 14 | 18 | 28 | -1 | -11 | -86 | -5 | 0 | 0 | 0 | 1 | 8 | |
| A.E. (mm) | 0 | 0 | 0 | 46 | 64 | 131 | 28 | 48 | 3 | 5 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 117 | 56 | 61 | 37 | 0 | 0 | 271 |

YEAR: 1975

| | J | F | M | A | M | J | J | A | S | O | N | D | ANNUAL TOTAL |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| MEAN AIR TEMP. (T°C) | - 8.4 | -14.2 | - 5.9 | -0.5 | 10.9 | 15.1 | 21.7 | 16.2 | 13.1 | 6.5 | - 2.4 | - 7.6 | |
| HEAT INDEX (I) | 0 | 0 | 0 | 0 | 3.25 | 5.33 | 9.23 | 5.93 | 4.30 | 1.49 | 0 | 0 | 29.53 |
| UNADJUSTED P.E. | 0 | 0 | 0 | 0 | 57 | 79 | 111 | 85 | 69 | 30 | 0 | 0 | |
| P.E. (mm) | 0 | 0 | 0 | 0 | 76 | 107 | 152 | 106 | 73 | 31 | 0 | 0 | 545 |
| P. (mm) | 8 | 21 | 37 | 53 | 87 | 70 | 37 | 34 | 23 | 14 | 25 | 26 | 435 |
| ST (12mm) | 12 | 12 | 12 | 12 | 12 | 0 | 0 | 0 | 0 | 0 | 12 | 12 | |
| ΔST. (mm) | 3 | 0 | 0 | 0 | 0 | -12 | 0 | 0 | 0 | 0 | 12 | 0 | |
| A.E. (mm) | 0 | 0 | 0 | 0 | 76 | 82 | 37 | 34 | 23 | 14 | 0 | 0 | |
| SURPLUS (mm) | 5 | 21 | 37 | 53 | 11 | 0 | 0 | 0 | 0 | 0 | -14 | 26 | 127 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 25 | 115 | 72 | 50 | 17 | 0 | 0 | 279 |
| ST (50mm) | 17 | 38 | 50 | 50 | 50 | 13 | 0 | 0 | 0 | 0 | 25 | 49 | |
| ΔST. (mm) | 8 | 21 | 12 | 0 | 0 | -37 | -13 | 0 | 0 | 0 | 25 | 24 | |
| A.E. (mm) | 0 | 0 | 0 | 0 | 76 | 107 | 50 | 34 | 23 | 14 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 25 | 53 | 11 | 0 | 0 | 0 | 0 | 0 | 0 | 2 | 89 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 102 | 72 | 50 | 17 | 0 | 0 | 241 |
| ST (100mm) | 17 | 38 | 75 | 100 | 100 | 63 | 0 | 0 | 0 | 0 | 25 | 51 | |
| ΔST. (mm) | 8 | 21 | 37 | 25 | 0 | -37 | -63 | 0 | 0 | 0 | 25 | 26 | |
| A.E. (mm) | 0 | 0 | 0 | 0 | 76 | 107 | 100 | 34 | 23 | 14 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 28 | 11 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 39 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 52 | 72 | 50 | 17 | 0 | 0 | 191 |
| ST (150mm) | 17 | 38 | 75 | 128 | 139 | 102 | 0 | 0 | 0 | 0 | 25 | 51 | |
| ΔST. (mm) | 8 | 21 | 37 | 53 | 11 | -37 | -102 | 0 | 0 | 0 | 25 | 26 | |
| A.E. (mm) | 0 | 0 | 0 | 0 | 76 | 107 | 139 | 34 | 23 | 14 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 13 | 72 | 50 | 17 | 0 | 0 | 152 |
| ST (250mm) | 17 | 38 | 75 | 128 | 139 | 102 | 0 | 0 | 0 | 0 | 25 | 51 | |
| ΔST. (mm) | 8 | 21 | 37 | 53 | 11 | -37 | -102 | 0 | 0 | 0 | 25 | 26 | |
| A.E. (mm) | 0 | 0 | 0 | 0 | 76 | 107 | 139 | 34 | 23 | 14 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 13 | 72 | 50 | 17 | 0 | 0 | 152 |

YEAR: 1976

| | J | F | M | A | M | J | J | A | S | O | N | D | ANNUAL TOTAL |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| MEAN AIR TEMP. (T°C) | - 7.4 | - 2.1 | - 2.7 | 8.7 | 14.2 | 15.7 | 20.0 | 19.9 | 15.7 | 5.9 | - 1.3 | - 4.6 | |
| HEAT INDEX (I) | 0 | 0 | 0 | 2.31 | 4.86 | 5.38 | 8.16 | 8.10 | 5.65 | 1.29 | 0 | 0 | 35.75 |
| UNADJUSTED P.E. | 0 | 0 | 0 | 41 | 70 | 75 | 100 | 100 | 77 | 27 | 0 | 0 | |
| P.E. (mm) | 0 | 0 | 0 | 47 | 93 | 102 | 137 | 125 | 82 | 25 | 0 | 0 | 611 |
| P. (mm) | 3 | 9 | 21 | 27 | 23 | 59 | 87 | 41 | 5 | 7 | 14 | 5 | 301 |
| ST (12mm) | 12 | 12 | 12 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 12 | 12 | |
| ΔST. (mm) | 0 | 0 | 0 | -12 | 0 | 0 | 0 | 0 | 0 | 0 | 12 | 0 | |
| A.E. (mm) | 0 | 0 | 0 | 39 | 23 | 59 | 87 | 41 | 5 | 7 | 0 | 0 | |
| SURPLUS (mm) | 3 | 9 | 21 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 2 | 5 | 21 |
| DEFICIT (mm) | 0 | 0 | 0 | 8 | 70 | 43 | 50 | 84 | 77 | 18 | 0 | 0 | 350 |
| ST (50mm) | 50 | 50 | 50 | 30 | 0 | 0 | 0 | 0 | 0 | 0 | 14 | 9 | |
| ΔST. (mm) | 0 | 0 | 0 | -20 | -30 | 0 | 0 | 0 | 0 | 0 | 14 | 5 | |
| A.E. (mm) | 0 | 0 | 0 | 47 | 53 | 59 | 87 | 41 | 5 | 7 | 0 | 0 | |
| SURPLUS (mm) | 3 | 9 | 21 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 21 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 40 | 43 | 50 | 84 | 77 | 18 | 0 | 0 | 312 |
| ST (100mm) | 55 | 64 | 85 | 65 | 0 | 0 | 0 | 0 | 0 | 0 | 14 | 19 | |
| ΔST. (mm) | 3 | 9 | 21 | -20 | -65 | 0 | 0 | 0 | 0 | 0 | 14 | 5 | |
| A.E. (mm) | 0 | 0 | 0 | 47 | 88 | 59 | 87 | 41 | 5 | 7 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 5 | 43 | 50 | 84 | 77 | 18 | 0 | 0 | 277 |
| ST (150mm) | 55 | 64 | 85 | 65 | 0 | 0 | 0 | 0 | 0 | 0 | 14 | 19 | |
| ΔST. (mm) | 3 | 9 | 21 | -20 | -65 | 0 | 0 | 0 | 0 | 0 | 14 | 5 | |
| A.E. (mm) | 0 | 0 | 0 | 47 | 88 | 59 | 87 | 41 | 5 | 7 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 5 | 43 | 50 | 84 | 77 | 18 | 0 | 0 | 277 |
| ST (250mm) | 56 | 64 | 85 | 65 | 0 | 0 | 0 | 0 | 0 | 0 | 14 | 19 | |
| ΔST. (mm) | 3 | 9 | 21 | -20 | -65 | 0 | 0 | 0 | 0 | 0 | 14 | 5 | |
| A.E. (mm) | 0 | 0 | 0 | 47 | 88 | 59 | 87 | 41 | 5 | 7 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 5 | 43 | 50 | 84 | 77 | 18 | 0 | 0 | 277 |

YEAR: 1977

| | J | F | M | A | M | J | J | A | S | O | N | D | ANNUAL TOTAL |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| MEAN AIR TEMP. (T°C) | -12.0 | 0.9 | 1.4 | 9.3 | 12.9 | 18.5 | 19.1 | 16.0 | 12.0 | 7.8 | - 3.5 | -12.7 | |
| HEAT INDEX (I) | 0 | 0.07 | 0.15 | 2.56 | 4.20 | 7.31 | 7.61 | 5.82 | 3.76 | 1.96 | 0 | 0 | 33.44 |
| UNADJUSTED P.E. | 0 | 4 | 6 | 46 | 64 | 94 | 97 | 80 | 60 | 35 | 0 | 0 | |
| P.E. (mm) | 0 | 3 | 6 | 52 | 85 | 128 | 123 | 100 | 64 | 35 | 0 | 0 | 596 |
| P. (mm) | 18 | 1 | 6 | 7 | 81 | 19 | 41 | 40 | 39 | 2 | 18 | 31 | 303 |
| ST (12mm) | 12 | 10 | 10 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 12 | 12 | |
| ΔST. (mm) | 0 | -2 | 0 | -10 | 0 | 0 | 0 | 0 | 0 | 0 | 12 | 0 | |
| A.E. (mm) | 0 | 3 | 6 | 17 | 81 | 19 | 41 | 40 | 39 | 2 | 0 | 0 | |
| SURPLUS (mm) | 18 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 6 | 31 | 25 |
| DEFICIT (mm) | 0 | 0 | 0 | 35 | 4 | 109 | 82 | 60 | 25 | 33 | 0 | 0 | 348 |
| ST (50mm) | 37 | 35 | 35 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 18 | 49 | |
| ΔST. (mm) | 18 | -2 | 0 | -35 | 0 | 0 | 0 | 0 | 0 | 0 | 18 | 31 | |
| A.E. (mm) | 0 | 3 | 6 | 42 | 81 | 19 | 41 | 40 | 39 | 2 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | - 0 | 10 | 4 | 109 | 82 | 60 | 25 | 33 | 0 | 0 | 323 |
| ST (100mm) | 37 | 35 | 35 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 18 | 49 | |
| ΔST. (mm) | 18 | -2 | 0 | -35 | 0 | 0 | 0 | 0 | 0 | 0 | 18 | 31 | |
| A.E. (mm) | 0 | 3 | 6 | 42 | 81 | 19 | 41 | 40 | 39 | 2 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 10 | 4 | 109 | 82 | 60 | 25 | 33 | 0 | 0 | 323 |
| ST (150mm) | 37 | 35 | 35 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 18 | 49 | |
| ΔST. (mm) | 18 | -2 | 0 | -35 | 0 | 0 | 0 | 0 | 0 | 0 | 18 | 31 | |
| A.E. (mm) | 0 | 3 | 6 | 42 | 81 | 19 | 41 | 40 | 39 | 2 | 0 | 0 | 0 |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 10 | 4 | 109 | 82 | 60 | 25 | 33 | 0 | 0 | 33 |
| ST (250mm) | 37 | 35 | 35 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 18 | 49 | |
| ΔST. (mm) | 18 | -2 | 0 | -35 | 0 | 0 | 0 | 0 | 0 | 0 | 18 | 31 | |
| A.E. (mm) | 0 | 3 | 6 | 42 | 81 | 19 | 41 | 40 | 39 | 2 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | -0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | |
| DEFICIT (mm) | 0 | 0 | 0 | 10 | 4 | 109 | 82 | 60 | 25 | 33 | 0 | 0 | 323 |

YEAR: 1978

| | J | F | M | A | M | J | J | A | S | O | N | D | ANNUAL TOTAL |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| MEAN AIR TEMP. (T°C) | -16.9 | -13.1 | - 2.4 | 6.3 | 12.1 | 17.1 | 19.5 | 17.6 | 14.0 | 8.5 | - 6.0 | -10.8 | |
| HEAT INDEX (I) | 0 | 0 | 0 | 1.42 | 3.81 | 6.44 | 7.86 | 6.72 | 4.75 | 2.22 | 0 | 0 | 33.22 |
| UNADJUSTED P.E. | 0 | 0 | 0 | 31 | 61 | 86 | 96 | 89 | 79 | 42 | 0 | 0 | |
| P.E. (mm) | 0 | 0 | 0 | 36 | 81 | 117 | 134 | 111 | 74 | 39 | 0 | 0 | 592 |
| P. (mm) | 19 | 16 | 9 | 72 | 55 | 53 | 45 | 34 | 93 | 31 | 23 | 9 | 459 |
| ST (12mm) | 12 | 12 | 12 | 12 | 0 | 0 | 0 | 0 | 12 | 4 | 12 | 12 | |
| ΔST. (mm) | 0 | 0 | 0 | 0 | -12 | 0 | 0 | 0 | 12 | -8 | 8 | 0 | |
| A.E. (mm) | 0 | 0 | 0 | 36 | 67 | 53 | 45 | 34 | 74 | 39 | 0 | 0 | |
| SURPLUS (mm) | 19 | 16 | 9 | 36 | 0 | 0 | 0 | 0 | 7 | 0 | 15 | 9 | 124 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 14 | 64 | 89 | 77 | 0 | 0 | 0 | 0 | 244 |
| ST (50mm) | 50 | 50 | 50 | 50 | 14 | 0 | 0 | 0 | 19 | 11 | 34 | 43 | |
| ΔST. (mm) | 1 | 0 | 0 | 0 | -26 | -14 | 0 | 0 | 19 | -8 | 23 | 9 | |
| A.E. (mm) | 0 | 0 | 0 | 36 | 81 | 67 | 45 | 34 | 74 | 39 | 0 | 0 | |
| SURPLUS (mm) | 18 | 16 | 9 | 36 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 79 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 50 | 89 | 77 | 0 | 0 | 0 | 0 | 216 |
| ST (100mm) | 68 | 84 | 93 | 100 | 74 | 10 | 0 | 0 | 19 | 11 | 34 | 43 | |
| ΔST. (mm) | 19 | 16 | 9 | 7 | -26 | -64 | -10 | 0 | 19 | -8 | 23 | 9 | |
| A.E. (mm) | 0 | 0 | 0 | 36 | 81 | 117 | 55 | 34 | 74 | 39 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 29 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 29 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 79 | 77 | 0 | 0 | 0 | 0 | 156 |
| ST (150mm) | 68 | 84 | 93 | 129 | 103 | 39 | 0 | 0 | 19 | 11 | 34 | 43 | |
| ΔST. (mm) | 19 | 16 | 9 | 36 | -26 | -64 | -39 | 0 | 19 | -8 | 23 | 9 | |
| A.E. (mm) | 0 | 0 | 0 | 36 | 81 | 117 | 84 | 34 | 74 | 39 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 50 | 77 | 0 | 0 | 0 | 0 | 127 |
| ST (250mm) | 68 | 84 | 93 | 129 | 103 | 39 | 0 | 0 | 19 | 11 | 34 | 43 | |
| ΔST. (mm) | 19 | 16 | 9 | 36 | -26 | -64 | -39 | 0 | 19 | -8 | 23 | 9 | |
| A.E. (mm) | 0 | 0 | 0 | 36 | 81 | 117 | 84 | 34 | 74 | 39 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 50 | 77 | 0 | 0 | 0 | 0 | 127 |

YEAR: 1979

| | J | F | M | A | M | J | J | A | S | O | N | D | ANNUAL TOTAL |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| MEAN AIR TEMP. (T°C) | -18.2 | -16.7 | 0.4 | 3.4 | 10.6 | 16.9 | 20.4 | 20.1 | 16.1 | 9.3 | - 1.3 | - 2.0 | |
| HEAT INDEX (I) | 0 | 0 | 0.02 | 0.56 | 3.12 | 6.32 | 8.41 | 8.22 | 5.87 | 2.56 | 0 | 0 | 35.08 |
| UNADJUSTED P.E. | 0 | 0 | 0 | 15 | 51 | 84 | 103 | 101 | 80 | 44 | 0 | 0 | |
| P.E. (mm) | 0 | 0 | 0 | 17 | 68 | 114 | 141 | 126 | 85 | 40 | 0 | 0 | 591 |
| P. (mm) | 6 | 16 | 8 | 26 | 46 | 76 | 32 | 20 | 15 | 18 | 7 | 11 | 281 |
| ST (12mm) | 12 | 12 | 12 | 12 | 0 | 0 | 0 | 0 | 0 | 0 | 7 | 12 | |
| AST. (mm) | 0 | 0 | 0 | 0 | -12 | 0 | 0 | 0 | 0 | 0 | 7 | 5 | |
| A.E. (mm) | 0 | 0 | 0 | 17 | 58 | 76 | 32 | 20 | 15 | 18 | 0 | 0 | |
| SURPLUS (mm) | 6 | 16 | 8 | 9 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 6 | 43 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 10 | 38 | 109 | 106 | 70 | 22 | 0 | 0 | 355 |
| ST (50mm) | 49 | 50 | 50 | 50 | 28 | 0 | 0 | 0 | 0 | 0 | 7 | 18 | |
| AST. (mm) | 6 | 1 | 0 | 0 | -22 | -28 | 0 | 0 | 0 | 0 | 7 | 11 | |
| A.E. (mm) | 0 | 0 | 0 | 17 | 68 | 104 | 32 | 20 | 15 | 18 | 0 | 0 | |
| SURPLUS (mm) | 0 | 15 | 8 | 9 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 32 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 10 | 109 | 106 | 70 | 22 | 0 | 0 | 318 |
| ST (100mm) | 49 | 65 | 73 | 82 | 60 | 22 | 0 | 0 | 0 | 0 | 7 | 18 | |
| AST. (mm) | 6 | 16 | 8 | 9 | -22 | -38 | -22 | 0 | 0 | 0 | 7 | 11 | |
| A.E. (mm) | 0 | 0 | 0 | 17 | 68 | 114 | 54 | 20 | 15 | 18 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 87 | 106 | 70 | 22 | 0 | 0 | 285 |
| ST (150mm) | 49 | 65 | 73 | 82 | 60 | 22 | 0 | 0 | 0 | 0 | 7 | 18 | |
| AST. (mm) | 6 | 16 | 8 | 9 | -22 | -38 | -22 | 0 | - 0 | - 0 | 7 | 11 | |
| A.E. (mm) | 0 | 0 | 0 | 17 | 68 | 114 | 54 | 20 | 15 | 18 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 87 | 106 | 70 | 22 | 0 | 0 | 285 |
| ST (250mm) | 49 | 65 | 73 | 82 | 60 | 22 | 0 | 0 | 0 | 0 | 7 | 18 | |
| AST. (mm) | 6 | 16 | 8 | 9 | -22 | -38 | -22 | 0 | 0 | 0 | 7 | 11 | |
| A.E. (mm) | 0 | 0 | 0 | 17 | 68 | 114 | 54 | 20 | 15 | 18 | 0 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 87 | 106 | 70 | 22 | 0 | 0 | 285 |

YEAR: 1980

| | J | F | M | A | M | J | J | A | S | O | N | D | ANNUAL TOTAL |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| MEAN AIR TEMP. (T°C) | -12.5 | - 7.0 | - 2.8 | 11.2 | 14.7 | 17.0 | 19.4 | 16.2 | 13.3 | 8.7 | 2.6 | - 9.2 | |
| HEAT INDEX (I) | 0 | 0 | 0 | 3.39 | 5.12 | 6.38 | 7.79 | 5.93 | 4.40 | 2.12 | 0.37 | 0 | 35.50 |
| UNADJUSTED P.E. | 0 | 0 | 0 | 54 | 73 | 84 | 96 | 80 | 65 | 38 | 11 | 0 | |
| P.E. (mm) | 0 | 0 | 0 | 62 | 97 | 114 | 132 | 100 | 69 | 36 | 8 | 0 | 618 |
| P. (mm) | 23 | 8 | 14 | 11 | 42 | 65 | 28 | 36 | 23 | 38 | 5 | 23 | 316 |
| ST (12mm) | 12 | 12 | 12 | 0 | 0 | 0 | 0 | 0 | 0 | 2 | 0 | 12 | |
| AST. (mm) | 0 | 0 | 0 | -12 | 0 | 0 | 0 | 0 | 0 | 2 | -2 | 12 | |
| A.E. (mm) | 0 | 0 | 0 | 23 | 42 | 65 | 28 | 36 | 23 | 36 | 7 | 0 | |
| SURPLUS (mm) | 23 | 8 | 14 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 11 | 51 |
| DEFICIT (mm) | 0 | 0 | 0 | 39 | 55 | 49 | 104 | 64 | 46 | 0 | 1 | 0 | 358 |
| ST (50mm) | 41 | 49 | 50 | 0 | 0 | 0 | 0 | 0 | 0 | 2 | 0 | 23 | |
| AST. (mm) | 23 | 8 | 1 | -50 | 0 | 0 | 0 | 0 | 0 | 2 | -2 | 23 | |
| A.E. (mm) | 0 | 0 | 0 | 61 | 42 | 65 | 28 | 36 | 23 | 36 | 7 | 0 | |
| SURPLUS (mm) | 0 | 0 | 13 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 13 |
| DEFICIT (mm) | 0 | 0 | 0 | 1 | 55 | 49 | 104 | 64 | 46 | 0 | 1 | 0 | 320 |
| ST (100mm) | 41 | 49 | 63 | 12 | 0 | 0 | 0 | 0 | 0 | 2 | 0 | 23 | |
| AST. (mm) | 23 | 8 | 14 | -51 | -12 | 0 | 0 | 0 | 0 | 2 | -2 | 23 | |
| A.E. (mm) | 0 | 0 | 0 | 62 | 54 | 65 | 28 | 36 | 23 | 36 | 7 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 43 | 49 | 104 | 64 | 46 | 0 | 1 | 0 | 307 |
| ST (150mm) | 41 | 49 | 63 | 12 | 0 | 0 | 0 | 0 | 0 | 2 | 0 | 23 | |
| AST. (mm) | 23 | 8 | 14 | -51 | -12 | 0 | 0 | 0 | 0 | 2 | -2 | 23 | |
| A.E. (mm) | 0 | 0 | 0 | 62 | 54 | 65 | 28 | 36 | 23 | 36 | 7 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 43 | 49 | 104 | 64 | 46 | 0 | 1 | 0 | 307 |
| ST (250mm) | 41 | 49 | 63 | 12 | 0 | 0 | 0 | 0 | 0 | 2 | 0 | 23 | |
| AST. (mm) | 23 | 8 | 14 | -51 | -12 | 0 | 0 | 0 | 0 | 2 | -2 | 23 | |
| A.E. (mm) | 0 | 0 | 0 | 62 | 54 | 65 | 28 | 36 | 23 | 36 | 7 | 0 | |
| SURPLUS (mm) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| DEFICIT (mm) | 0 | 0 | 0 | 0 | 43 | 49 | 104 | 64 | 46 | 0 | 1 | 0 | 307 |

YEAR: AVERAGE

| | J | F | M | A | M | J | J | A | S | O | N | D | ANNUAL TOTAL |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| MEAN AIR TEMP. (T°C) | | | | | | | | | | | | | |
| HEAT INDEX (I) | | | | | | | | | | | | | |
| UNADJUSTED P.E. | | | | | | | | | | | | | |
| P.E. (mm) | | | | | | | | | | | | 23.3 | 602.0 |
| P. (mm) | | | | | | | | | | | | 12.9 | 327.6 |
| ST (12mm) | | | | | | | | | | | | | |
| AST. (mm) | | | | | | | | | | | | | |
| A.E. (mm) | | | | | | | | | | | | | |
| SURPLUS (mm) | | | | | | | | | | | | 2.87 | 72.6 |
| DEFICIT (mm) | | | | | | | | | | | | 13.3 | 336.2 |
| ST (50mm) | | | | | | | | | | | | | |
| AST. (mm) | | | | | | | | | | | | | |
| A.E. (mm) | | | | | | | | | | | | | |
| SURPLUS (mm) | | | | | | | | | | | | 1.78 | 35.3 |
| DEFICIT (mm) | | | | | | | | | | | | 11.9 | 302.0 |
| ST (100mm) | | | | | | | | | | | | | |
| AST. (mm) | | | | | | | | | | | | | |
| A.E. (mm) | | | | | | | | | | | | | |
| SURPLUS (mm) | | | | | | | | | | | | 0.27 | 6.8 |
| DEFICIT (mm) | | | | | | | | | | | | 10.8 | 273.2 |
| ST (150mm) | | | | | | | | | | | | | |
| AST. (mm) | | | | | | | | | | | | | |
| A.E. (mm) | | | | | | | | | | | | | |
| SURPLUS (mm) | | | | | | | | | | | | 0 | 0 |
| DEFICIT (mm) | | | | | | | | | | | | 10.5 | 266.4 |
| ST (250mm) | | | | | | | | | | | | | |
| AST. (mm) | | | | | | | | | | | | | |
| A.E. (mm) | | | | | | | | | | | | | |
| SURPLUS (mm) | | | | | | | | | | | | 0 | 0 |
| DEFICIT (mm) | | | | | | | | | | | | 10.5 | 266.4 |